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CANONS OF LANDSCAPE EVOLUTION

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Abstract

The manner in which epigene landscapes evolve is examined and discussed. Slope flattening as a general process of landscape evolution is rejected, and with it Davis' concept of the peneplain. Landscape evolution by scarp retreat and pedimentation is accepted, and several of its consequences are examined.

The opinions derived are expressed at the end of the paper as a series of canons of landscape development

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Historical Approach

The scientific study of landscape began in the latter half of the nineteenth century and is associated especially with the names of J. W. Powell, G. K. Gilbert, and W. M. Davis in America and Albrecht Penck in Europe. The work of the American trio has always overshadowed in the public regard that of Penck senior, and modern thought still follows closely the pronouncements and opinions of the last of the trio—William Morris Davis. Davis had a remarkably analytical mind, he reasoned lucidly and logically; moreover he wrote with facility, grace, and copiousness, so that his ideas were easily and pleasantly assimilated, became widely disseminated, and carried conviction. He was an inspired teacher, and his ideas found a ready audience and ready acceptance.

Particularly was this so with his concept of

the "Cycle of Erosion" under which a landscape uplifted and subjected to the forces of subaerial erosion was first vigorously dissected by the deeply cut valleys of the rejuvenated river systems, with consequent increase of relief, and later reduced by weathering and lowering of the interfluves to a lowland of faint relief which he termed a *peneplain*. The type areas cited for these changes were the areas in which Davis worked in his early professional life, chiefly from Pennsylvania to New England; and thus something of a parochial standard of comparison was instituted for other regions of the globe. Landscapes which did not conform to this standard, such as desert landscapes, were regarded as "abnormal", "accidental", or exceptional.

In his later years, Davis wrote many fine papers, yet none to my mind was more truly philosophical nor more important than his contribution of 1930 wherein the earlier view was revised and many true homologies demonstrated between the landscapes of arid and of humid regions. This harmony calls for revision of a number of the earlier concepts, and particularly the standard of "normality" in landscape types, but the necessary revisions have not been undertaken, with any degree of completeness, either by Davis or by others.

The review of what Davis called "the Cycle of Erosion" is thus broadly the theme of the present paper; but before proceeding with our argument we shall adduce a little more of the historical background.

The cyclic concept did not originate with Davis; it was stated first in Powell's *Exploration of the Colorado River of the West* (1875), in the following terms (we quote from *Geographical Essays*)

> " 'aerial forces carried away 10,000 feet of rocks by a process slow yet unrelenting until the sea again rolled over the land' and the evenly denuded surface is referred to as 'the record of a long time when the region was land' ".

Powell recognized very clearly that the later stages of reduction in the landscape would be very slow:

> "The degradation of the last few inches of a broad area of land above the level of the sea would require a longer time than all the thousands of feet that might have been above it, so far as this degradation depends on mechanical process—that is by driving or flotation; but here the disintegration by solution and the transportation of material by the agency of fluidity come in to assist the slow processes of mechanical disintegration, and finally perform the chief part of the task".

The role of solution as a significant agent of reduction in the later stages of landscape development is open to question. Many senile land surfaces bear a residuum of insoluble matter as a subsurface crust—laterite, silcrete, caliche are examples—but all these hinder degradation in a marked degree, and it is doubtful whether solution is of any real significance in the final reduction of noncalcareous terrains. We shall ignore it in our further discussion.

Davis' specific addition to Powell's theory was to visualize and describe the landforms that, under the action of erosional processes, were evolving in the landscape toward a final stage.

The contribution of Walther Penck (1924), in which the forms of hillslopes are related to continuing earth movements, has not found any enduring acceptance. Nevertheless, Penck junior understood certain classes of landforms (*e.g.*, hillslopes) better than Davis did, and several of his viewpoints such as the parallel retreat of scarps and the difference between senile and initial landscapes are real in nature and make an advance on the earlier Davisian technique.

Landscape studies have now advanced to the stage where statistical analysis is being applied to certain classes of data (Strahler, 1950b). Used critically, so that it does not lead to erroneous conclusions that become established in the minds of scientists, this method should tend to much more exact definition of landforms, processes, and principles. Undoubtedly, if the basic data can be supplied from observation and measurement of landscapes, the method is ideal for disentangling the complexities of forms and processes governed respectively by many factors or variables. What other equally satisfactory approach could there be, for instance, to the complexities of stream flow and grade?

ACKNOWLEDGMENT

The script has benefited from a critical reading by my colleague Dr. T. J. D. Fair, whose contributions are far more important

than the few footnotes to his name indicate. Several phrases and sentences have been reframed at his suggestion, and to him I am indebted for the canons numbered (1), (12), and (22). Several of the opinions expressed have also been derived from his careful field studies and published works. All his assistance is gratefully acknowledged.

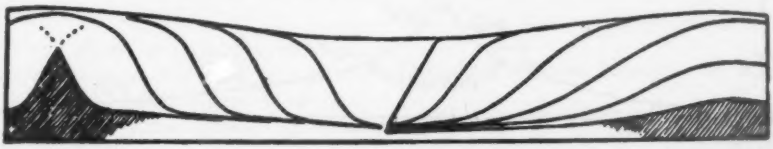

FIGURE 1.—CONTRASTED SLOPE PROFILES

Widening of valleys and change in slope profile during the cycle. Right, youth to old age, according to W. M. Davis. Left, with parallel scarp retreat, according to W. Penck. (From Davis, 1930)

STATEMENT OF THE PROBLEM

Davis' original conception of landscape evolution under the subaerial processes of rainfall, running water, and weathering has often been summed up in a series of progressive sections in which an uplifted landscape is depicted as first dissected under stream incision with the production of narrow valleys. After the streams have attained grade (defined as the condition in which all major irregularities have been eliminated from the thalweg), the rate of river incision is reduced to negligible proportions, so that the valley bottoms are lowered very little during the remainder of the cycle. According to Davis, the valley sides are then reduced under weathering and surface creep, and perhaps wash, to ever flatter and flatter angles, until they meet upon the interfluves and destroy the last remnants of the initial surface. As the slopes continue to flatten, the interfluves are lowered more rapidly than the river beds until only "a lowland of faint relief remains". To this ultimate landform Davis assigned the title of *peneplain*. The concept is illustrated by Davis' own diagram which is here reproduced (Fig. 1). The progressive lowering of the interfluves under weathering is a vital concept in the "Normal Cycle" as visualized by Powell, by Davis, and by Cotton.

The Davisian concept has not passed without challenge, notably in his own Harvard University, where the strong school of Kirk Bryan accepted parallel retreat of scarps (*q.v.*). Some authorities have indeed rejected the cyclic concept altogether (Penck, 1924); others, with whom we align ourselves, have accepted the general concept of a cycle of landforms developed under erosion, while considering that the detailed forms and sequences depart considerably from those visualized and adduced by Davis. The essential differences of viewpoint lie in the interpretation of slope forms and the manner of hillslope development.

In summary, this opinion, after beginning the cycle of erosion with stream incision into an uplifted land surface exactly as in the Davisian model, would place limits to the amount of hillslope flattening, regarding the slopes as attaining a stable gradient (defined for local conditions) after which the upper parts of the slopes retreat parallel to themselves. A well-known example is due to Kirk Bryan, 1922, p. 42–46. At the foot of the slope is left a *pediment* sloping gently down to the river. This pediment is concave in profile, so that, when the interfluves are consumed by parallel scarp retreat, the opposing pediments from adjacent valleys would meet without forming a broad convexity across the interfluve (Figs. 1, 2).

The ultimate landscape under this philosophy, being composed of many coalescing pediments, is termed a *pediplain*, and it is distinguishable at sight from the Davisian peneplain by the multi-concave instead of multi-convex nature of its surface and the presence of steep-sided rather than gentle residuals. The fundamental difference between the two is, of course, not merely a difference of surface form—which indeed may be dubious, for concavities might occur in relation to peneplains—but of different history and mode of development.[1]

[1] In discussion Dr. T. J. Fair has commented upon this point: "Though Davis neither measured slopes nor, apparently, appreciated the processes

The two concepts of an erosion cycle are largely exclusive. Over any given area, one can be true, but not the other. In the viewpoint of many geologists, the so-called "Normal Cycle of Erosion" appears under humid climates, and the Pediplanation cycle under semiarid and arid climates. The viewpoint is not without absurdity. One can understand that a characteristic cycle and set of landforms should be generated in frigid zones where water is frozen into ice and the whole mechanism of abrasion alters, where cirques are formed by thaw and freeze under specific conditions (D. W. Johnson), and where roches moutonnées are overridden by masses of solid ice; but that mere differences in amount and incidence of rainfall, evaporation, and like factors should have such far-reaching effects as to result in two entirely distinct cycles requires further investigation. The primary agent moulding the landscape in both humid and arid cases is water flow, and this should produce comparable results in both types of region.

This, indeed, was realized by Davis who summarized his position (1930, p. 145) on landforms of humid and of arid regions:

"The comparison . . . will show that, while various contrasted features are truly enough unlike in certain respects, their unlikeness is rather a matter of degree than of kind. The apparently unlike features are really homologous; their resemblances, once recognized, are much more striking than their differences. The same may be said of the erosional and degradational processes by which apparently unlike forms are sculptured".

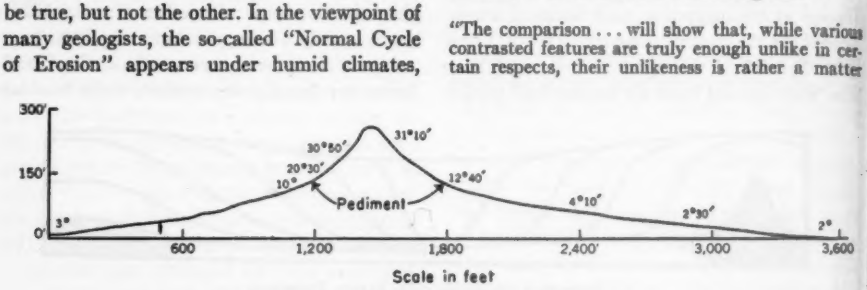

FIGURE 2.—SLOPES OF A NATAL MONADNOCK

Surveyed by T. J. D. Fair. On either side is a broad concave pediment, and the hillsides are steep (30°). Summit convexity is small. This type of monadnock agrees much better with Penck's doctrine of parallel scarp retreat than with Davis' flattening slope profiles and widespread convexity (Fig. 1).

He instituted comparisons between the processes of weathering both mechanical and chemical, soil creep and rill wash, stream pattern and action, the respective mantles of waste and types of hillslope (p. 147).

"As the various processes of arid and of humid erosion are thus seen to differ in degree and manner of development rather than in nature, and as their differences in degree and manner are wholly due to differences in their climates, so the forms produced by the two erosions may be shown to differ in the degree to which certain elements of form are developed rather than in the essential nature of the elements themselves".

Davis here stood upon the threshold, with the unification of epigene landscapes before him. But the shackles of his earlier thought, nearly 50 years upon him, were not to be broken so easily, and he never entirely accepted real identity of landscape evolution under humid and under arid influences, which is one of our present canons. Thus (1930, p. 147),

". . . although arid erosion resembles humid erosion in many respects the two kinds of erosion nevertheless differ so much in process and in product that they cannot be clearly understood if they are briefly brought together as examples of normal erosion, as has been done by several of the German writers. . . . The unlikenesses of the two erosions deserve explicit treatment".

operative upon them, he was nevertheless a master of landscape observation and I find it difficult to believe that he would see on his peneplains dominant convexities when, in reality the form was dominantly concave. I'd feel happier on this point if I could measure low divides in humid-temperate regions but, whatever the conclusion, I don't see that the possibility that peneplains and pediplains have more or less the same form invalidates the pediplanation concept one bit—for it is the process and not the end-form that is the crucial matter here. Can we say (Canon 37) the peneplain is 'an imaginary landform'? I am inclined to believe that Davis *really saw* pediplains, but called them peneplains, and that what is imaginary is not the peneplain itself but the peneplanation process. Did not Davis observe the features of the landscape truly but fail in their genetic analysis?"

Davis still held, moreover, to his earlier opinion that hillslopes flatten progressively under humid regimes (1930, p. 148):

"A striking difference between humid and arid (granitic) mountains is found in the decreasing angle of slope of the first as the erosion cycle advances, in contrast to the constant angle of slope in the second".

From this we dissent, and regard the manner of hillslope evolution as dependent upon factors such as bedrock, relief, vegetal cover, etc., rather than upon direct climatic influences. We dissent likewise from his separation of monadnocks from inselbergs. The latter are merely a special case of the former, characteristic of granitic and gneissic terrains. Monadnocks are not, as Davis claimed, "pale forms of weakening convexity and lessening slopes" but are truly concave in profile as indeed is the classic example, Mt. Monadnock itself. Davis' theoretical viewpoint of peneplanation obtrudes here and causes misinterpretation of visible landforms.

Still later (1932, p. 408), rejecting Penck's doctrine of parallel retreat of slopes, Davis asserts that:

"... the retreat of a valley side is usually accompanied by a decrease in the steepness of its slope as well as by the development of a convex profile at its top and a concave profile at its base whatever its original profile may have been. This is largely because a shoulder ... either angular or sharply rounded, at the top of a valley side yields more rapidly to the attack of weathering and creeping on its two faces than do the more nearly plane surfaces adjoining the shoulder; and at the same time the coarser detritus supplied by the steeper part of the slope and delivered to its base demands a stronger declivity there for its further downhill carriage by wash and creep than the same detritus will require when it has been reduced to finer texture as it advances. Hence even a square-shouldered top edge of a valley-side ... may soon wear back into a round-shouldered top and the ... round shoulder of short arc and rapid curvature will in later time change into a longer arc of gentler curvature. The concave basal curve is likewise enlarged as it retreats".

It all reads so logically, yet in my experience mapping major scarps and valley sides that have existed, some of them even from Mesozoic time (1944; 1947; 1951a), Penck was right about parallel scarp retreat, and Davis was wrong. Scarp retreat is normally more rapid than pronounced rounding. Otherwise these scarps would have been smoothed out long ago; on the contrary, they are still as steep as is consistent with the rock types of which they are composed (Pl. 1, fig. 1). The final court of appeal is the brutal facts of Nature, as displayed in landscape (Fig. 2).

The Nature of Hillslopes

Evolution of Hillslopes According to W. M. Davis

Davis considered that valley sides behaved rather as rill profiles of indefinite lateral extent, and applied to them the same concept of "grade" that he had used in the development of river profiles. Indeed, the rules governing valley side or hillslope development did not differ, in his view, from those governing flow, transport, and grade in a river system.

Thus Davis said (1909, p. 266) on the development of graded valley sides:

"When the migration of divides ceases in late maturity, and the valley floors of the adjusted streams are well graded, even far towards the headwaters, there is still to be completed another and perhaps even more remarkable sequence of systematic changes than any yet described: this is the development of graded waste slopes on the valley sides. It is briefly stated that valleys are eroded by their rivers, yet there is a vast amount of work performed in the erosion of valleys in which rivers have no part. It is true that rivers deepen the valleys in the youth and widen the valley floors during the maturity and old age of a cycle, and that they carry to the sea the waste denuded from the land; it is this work of transportation to the sea that is peculiarly the function of rivers, but the material to be transported is supplied chiefly by the action of the weather on the steeper inconsequent slopes and on the valley sides. The transportation of the weathered material from its source to the stream in the valley bottom is the work of various slow-acting processes, such as the surface-wash of rain, the action of ground water, changes of temperature, freezing and thawing, chemical disintegration and hydration, the growth of plant roots, the activities of burrowing animals. All these cause the weathered rock waste to wash and creep slowly downhill. ... In the first place a waste sheet moves fastest at the surface and slowest at the bottom, like a waterstream".

"In the second place ... waste sheets normally begin to establish a graded condition at their base and then extend it up the slope of the valley-side whose waste they 'drain' ". Then follows a most important passage (p. 268): "Just as graded rivers slowly degrade their courses after the period of maximum load is past, so graded waste sheets adopt gentler and gentler slopes when the upper ledges are consumed and coarse waste is no longer plentifully shed to the valley sides below. A changing ad-

justment of a most delicate kind is here discovered. When the graded slopes are first developed they are steep, and the waste that covers them is coarse and of moderate thickness; here the strong agencies of removal have all they can do do dispose of the plentiful supply of coarse waste from the strong ledges above, and the no less plentiful supply of waste that is weathered from the weaker rocks beneath the thin cover of detritus. In a more advanced stage of the cycle, the graded slopes are moderate, and the waste that covers them is of finer texture and greater depth than before; here the weakened agencies of removal are favored by the slower weathering of the rocks beneath the thickened waste cover, and by the greater refinement (reduction to finer texture) of the loose waste during its slow journey. In old age, when all the slopes are very gentle, the agencies of waste removal must everywhere be weak, and their equality with the processes of waste supply can be maintained only by the reduction of the latter to very low values. The waste sheet thus assumes great thickness—even fifty or a hundred feet—so that the progress of weathering is almost *nil*; at the same time the surface waste is reduced to extremely fine texture, so that some of its particles may be moved even on faint slopes. Hence the occurrence of deep soils is an essential feature of old age, just as the occurrence of bare ledges is of youth".

Again, of Old Age: "Maturity is passed and old age fully entered upon when the hilltops and the hillsides, as well as the valley floors, are graded".

It has been deemed necessary to quote fairly fully the older concept of hillslope evolution because I found on a recent tour of Europe and North America that many geologists in those regions still accepted it as "normal" or standard.

As late as 1950 in a study by Strahler, we encounter this same concept of flattening, graded slopes in an extended form. His equilibrium theory of erosional slopes (1950a) states that:

"slopes maintain an equilibrium angle proportional to the channel gradients of the drainage system and are so adjusted as to permit a steady state to be maintained by the process of erosion and transportation under prevailing conditions of climate, vegetation, soils, bedrock and initial relief or stage. Thus both slopes and streams are graded. . . . As the landmass is reduced both slopes and stream gradients are reduced, being slowly and continuously regraded to maintain approximate equilibrium. As the correlation of stream gradients with slopes suggests, the decline of stream gradient is accompanied by slope reduction. This concept of maintenance of a steady state by slow readjustment is essentially Davis's concept of landmass development in the normal cycle".

But all this fine Davisian armchair philosophizing bears scant semblance of reality to workers from other regions. Its relation to hillslopes and valley sides as observed in the field is often remote. Such features are infinitely more complex than Davis with his single criterion of "grade" would have us believe; and, while the general tenet of "grade" may be ideally true if no other factors existed, certain of Davis' conclusions are demonstrably untrue. We cite but three.

(1) Deep soils do not necessarily, or even characteristically, appear upon old-age surfaces. The plateau landscapes of the high African interior are much older, for instance, than any of the land surfaces of the eastern United States with which Davis was early familiar. Yet soils upon them are universally thin (often only a few inches) and poor in quality. Soils on landscapes a twentieth as old in Europe are deep and fertile. Landscapes of continental extent reverse the Davisian conclusion!

(2) The philosophy calls for progressively flatter slopes, measured from interfluve to stream bed, as the cycle proceeds (flattening of interfluves). While some slopes, or many, may be shown to have flattened in the course of their history, multitudes of hillslopes have clearly not done so. Frequently, in a homogeneous region, the maximum slopes measured, upon large and small hills alike, cluster closely about a mean, affording proof positive that, once a stable angle has been achieved, the slopes have not flattened any more, but have retreated parallel to themselves. Many ancient landscapes, moreover, bear steep-sided residuals (Pl. 3, figs. 1, 3). There have, in all these instances, clearly been no changes in the factors governing slope form, such as is implied under the concept of flattening.

And if further proof were needed, what of steep mountain scarps like the wall of the Drakensberg (Pl. 1, fig. 1), which are oriented at right angles to the flow of major river systems. These scarps, generated tectonically in the remote geologic past, in intimate relation to the initiation of new erosion cycles, have retained their steepness unimpaired despite retreat of scores of miles (Fig. 3). The Drakensberg happens to constitute a watershed, but other scarps of similar form run through Natal athwart the middle courses of the major rivers

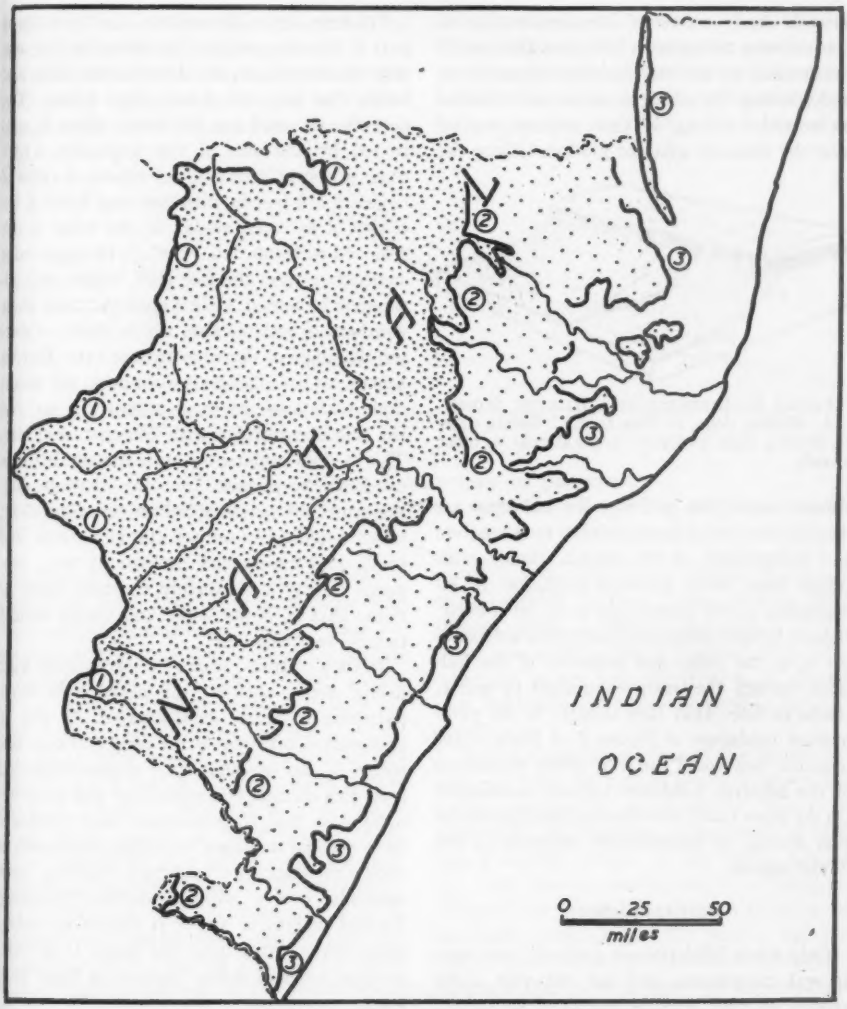

FIGURE 3.—CYCLIC EROSION SCARPS IN NATAL, SOUTH AFRICA

Showing diagrammatically the major river systems and the trends of the major cyclic erosion scarps (in each case over 1000 feet high). These scarps have retreated successively westward scores of miles from an original position near the present coast. The Drakensberg, the oldest, has retreated over 100 miles since the late Mesozoic and is still wall-like and over 4000 feet high (Pl. 1). The scarps are (1) Drakensberg, (2) Ixopo-Pietermaritzburg-Greytown scarp, (3) coastal-hinterland scarp.

(King, 1947) (Fig. 3). Similar features have been mapped or noted in other parts of the world. Even granting that a majority of these features have been tectonically induced, why have they retreated and why has their steepness not disappeared in the long ages since they were initiated?

(3) The development of a broad summit convexity upon mature and senile divides is a necessary corollary of the Davisian hypothesis. Such convexity is often absent from these situations, a point which we shall discuss under the heading of Peneplain or Pediplain.

The Equilibrium Theory of Slope Develop-

ment is surely countered by consideration of the processes acting upon hillslopes. Only while the streams are actively incising themselves or undercutting the adjacent spurs and hillslides on meander curves do they exercise control over the forms of adjacent hillslopes. Once the

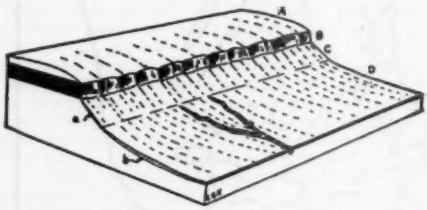

FIGURE 4.—ELEMENTS OF A HILLSIDE SLOPE
A. Waxing slope, B. Free face, C. Debris slope, D. Waning slope (pediment in the sketch). a. talus, b. soil.

streams cease this activity, the hillslopes are moulded by a set of forces peculiar to themselves and independent of the stream. Hence, with stable base level, hillslope evolution is independent of the later stages of the river cycle. Indeed, in these stages, the river cycle is dependent upon the forms and evolution of the hillsides through the waste—abundant or sparse, coarse or fine—that they supply. In the pedimented landscape of Figure 2 of Plate 1, for instance, how could the river affect the nature of the adjacent hillslopes without re-incision? On the other hand, the effects of hillslope evolution should be immediately reflected in the fluvial regime.

Hillslope Elements

Only when hillslopes are analyzed into their several components and the behavior under erosion of each part is examined separately can much progress be made in the study of hillslope evolution. This necessary analysis was undertaken by Wood (1942), and much of the modern study of cyclic erosion begins at this point.

Wood recognized that four distinct elements are sometimes present within the compass of a single hillside. From top to bottom (Fig. 4) these are: (1) the waxing slope (convex), (2) the free face (rock outcrop), (3) the talus or debris slope, (4) the waning slope (including pediments) (concave).

The free face is the outcrop, on the steepest part of the slope, of local bedrock. As the outcrop weathers back, the debris slides clear and builds the talus or debris slope below. The declivity assumed by the debris slope is controlled by the size of the fragments, which come to rest at their angle of repose. A talus of boulders 3–4 feet in diameter may have a declivity of as much as 35°. If the talus is not removed with sufficient rapidity, its upper edge encroaches upon the free face, which may ultimately be buried. Debris upon the talus slope weathers to smaller size, which tends to flatten the angle of slope toward the base. But, in general, as the hillslope is worn back, the debris slope maintains the same proportion, so that the hillslope retreats with constant declivity. This it does apparently even though little or no debris be present. In many districts, hillslopes exhibit remarkable likeness of gradient, which could be so only if the hills, large and small, had maintained throughout their erosional history a reasonably constant angle of slope; that is, if the slopes and scarps retreat parallel to themselves[2].

In such districts, where free faces or at least debris slopes abound, the absence of slope flattening is clearly demonstrated. As will be appreciated, hillslopes in the early and very late stages of the erosional cycle depart somewhat from this dictum, but experience and measurement show that, from the time major hillslopes have become stabilized until they lose much of their importance as landscape features, they normally evolve without decided flattening. Parallel retreat of slopes is dependent solely upon forces acting upon the slope. It is independent even of minor changes of local base level.

The agencies which effect the retreat of a hillside scarp are weathering, downhill movement of debris under gravity, sheet wash, and the attack of the gully heads of numerous small streams or rills which correspond, in a measure,

[2] For "stable angle of slope" there may be composite slopes where contrasted rock types appear in the valley sides—*e.g.*, Grand Canyon of Arizona. Under these circumstances minor free faces alternate with debris slopes. Fair has stated (1947) that it is the outcrops which govern the form of the slope as a whole, and that the slope will be less steep in maturity than with a single cap-rock (Fig. 5).

to the free face. Of these, the gully heads are usually by far the most important.

Free faces and debris slopes are usually best developed in regions of moderate to high relief. In regions of low relief, these two elements of hillside as the free face. As the edge of this zone is unbuttressed, the weathered mantle naturally moves off under soil creep. The curve is convex because more waste must pass a point lower down the slope in a unit of time than

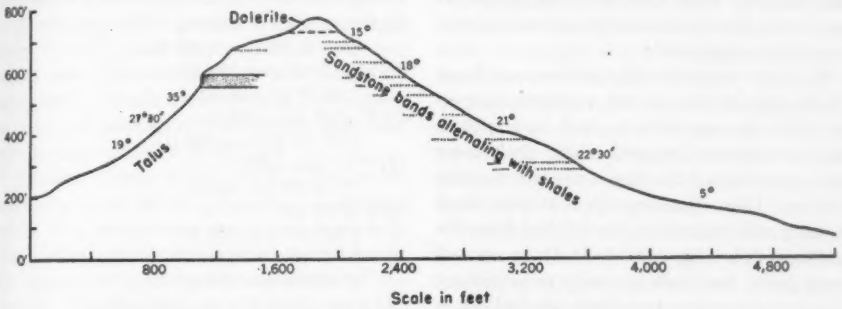

FIGURE 5.—EFFECT OF STRUCTURE ON HILLSLOPES

The single dominating sandstone band (left) generates a steeper slope than several minor bands (right). After Fair.

the hillside slope may be absent, and the waxing slope from above then makes conjunction with the waning slope from below. Under these circumstances, the slope development by parallel retreat is hindered or aborted. Here is an important principle! Large scarps normally exhibit considerable free faces and debris slopes, and hence such scarps undergo parallel retreat. So arises the curious fact, which I have noted in previous publications, that major cyclic erosion scarps in many continents retreat, apparently, at a sensibly uniform rate. This rate is between 1 foot in 150 years and 1 foot in 300 years. Smaller scarps and hillsides, however, may retreat with extreme slowness. Apparently, if the waxing and waning slopes meet so that the free face and debris slope are eliminated, parallel retreat of the scarp may be so slowed as almost to cease. The landscape of countries like the British Isles where concavo-convex slopes are widespread may thus be altering only with extreme slowness compared with landscapes possessing clear scarps with free faces and debris slopes. In this connection, the low relief of most of the British Isles is to be noted.

The waxing slope at the top of a hillside is convex in section. In simple instances it corresponds to the edge of the weathered zone on top of firm bedrock which crops out on the passes a point higher up. Convex slopes are thus moulded by soil creep (Davis, 1892; Gilbert, 1909). Lawson's explanation (1932) of the convexity as due to rainwash over saturated soil is not here admitted as a general case. All freely cut hydraulic profiles are concave.

Waning slopes at the base of the hillside exhibit in section the typical hydraulic curve of a surface modelled by water erosion. This is so whether the waning slope merges gently upward into a convex profile of the upper hillside (*e.g.*, the concavo-convex slopes of Southern England) or whether it abuts with more or less sharp angle against the base of the free face or debris slopes (*e.g.*, Southern Rhodesia). All gradations between these two extremes appear in nature.

Beneath the surface wash of the waning slope, either of two subsurface states may exist. First, the subjacent firm rock may pass upward through progressively weathered mantle into soil, in which case the concave surface profile is probably developed only in weathered materials and by surface wash; or, second, the solid rock is sharply truncated by an erosion bevel above which lies an accumulation of transported debris. This latter is the typical condition upon pediments. Of course, landforms that were originally cut-rock pediments may, if the cycle

of hillslope development is arrested, deteriorate into the first condition as the transported mantle is slowly removed and weathering proceeds beneath it into the bedrock. Some of the concavities of lower hillslope profiles of northern latitudes may thus have originated as pediments, though their origin as such may no longer be demonstrable.

Normally veneered with detritus, most if not all of which is transported, pediments are essentially cut-rock features, and bare rock is not uncommonly exposed where the veneer thins away toward the base of the commanding hillslope. Here often appears a distinct break in the profile separating the hillslope from the pediment (Pl. 1, fig. 2; Pl. 4, fig. 1). In areas of weak rocks, the break is usually smoothed out into an open curve, but where the bedrock is hard and there is little or no accumulated waste the junction between pediment and free face may be so sharply angular that a boot can scarcely be placed in it (Pl. 2, fig. 1). Sometimes, such abrupt transitions are sharpened by joining, but this is not necessary, and many of the boundaries intersect the joint systems at an angle.

Instances have also been quoted of nicks or inverted pediments. Bryan regarded any explanation of such nicks other than by weathering and rainwash as fantastic. Indeed, many may be shown to be located along semihorizontal joints. They are special features, and are not essential to pediments or pediment theory.

Occasionally, where the inner zone of the pediment is of bare rock or where the pediment extends as a rock fan into a scarp, it is convex over small areas, but such details, though not uncommon, merely represent incomplete stages in the production of the inner pediment.

Often the angular relation in the bedrock is masked by debris fallen or washed from above. Fair (1947, p. 109) has regarded the zone as one in which the coarse waste of the debris slope is broken down under weathering ready for transportation across the lower declivity of the pediment. "Thus the rate at which waste is comminuted determines in no small measure the sharpness of the angle between talus and pediment". Both Lawson and Bryan have also attributed the abrupt change of slope between hillside and pediment to a relatively sudden change from coarse debris on the hillside to finer debris on the pediment. But this can not be the whole truth for the change is even more abrupt, as we have seen, when no debris is present. The essence of the matter seems to be that there is a fundamental change of geomorphic process acting upon the steep and the gentle slopes respectively to mould them into their present forms. Gravity for instance ceases to be a potent factor, and the activity of running water changes.

TABLE 1.—TYPICAL GRADIENTS UPON PEDIMENTS*

(a) 3° to 5°50′ over 4700 ft:
5°50′ to 8°50′ over 200 ft.
(b) 2°50′–5°10′ over 2800 ft:
5°10′–9°20′ over 300 ft.
(c) 1°–4°50′ over 2600 ft:
4°50′–12°40′ over 500 ft.
(d) 2°–5° over 2250 ft:
5°–11°30′ over 500 ft.
(e) 4°–4°50′ over 2150 ft:
4°50′–13° over 650 ft.
(f) 1°10′–5° over 1350 ft:
5°–12°30′ over 300 ft.

* From Fair (1947, p. 110), who says: "Though it rises to a maximum angle of 13 degrees the pediment is essentially a slope of less than 5 degrees over by far the greater part of its (profile) length".

Pediments originate and grow by retreat of the hillside elements (free face and debris slope) above them. Thereafter, as their concave profiles guarantee, they are modelled under the influence of running water. The curve so generated is obviously a function of the age and width of the pediment (Table 1).

In certain regions of the globe—*e.g.*, the eastern and southern districts of Southern Rhodesia (King, 1951a)—pediments constitute 60 per cent or more of the landscape. Indeed, King (1949b) has regarded them as possibly the most important of all landforms. The pediment, he has there written, "is the fundamental form to which most, if not all, subaerial landscapes tend to be reduced [under erosion], the world over".

PEDIMENTATION

Much thought has been devoted to the mechanisms by which pediments originate and

grow, and a diversity of opinion has been expressed.

McGee (1897, p. 108), attributed the cutting of the wide cut-rock surfaces of the Sonora district, Mexico, to the action of sheet flood. But sheet floods of the magnitude described by him are rare, while pediments can be ubiquitous, so, while both may appear at their best under dry climates, there is not necessarily any genetic relation between them. Sheet floods may mould pediments that already exist, but how do they generate pediments?

Johnson (1931; 1932a; 1932b) laid great emphasis upon lateral planation by streams where they emerge from the mountain front to trim back the scarp and originate the pediment. He pointed to a number of rock fans at the mouths of valleys emerging through the scarp and considered that the scarp was trimmed back when the stream happened to flow along the extreme lateral radius of the fan and impinged against the scarp front. This action should develop a zig-zag outline for the scarp foot, re-entrants of pediment occurring up all the valleys of the scarp face with promontories being left between. Such an outline is, however, not characteristic of the scarps behind pediments; the scarps are straight or only gently sinuous. Likewise the inner edge of pediments should be markedly uneven in the horizontal sense, rising toward and into the valley mouths and falling to intermediate points between them as the successive rock fans are passed. While such features do occur on a minor scale, the inner edge of a normal pediment impresses an observer most because it is level over long distances. Moreover, stream-cut nicks at the foot of scarps are nonexistent.

For the modelling of pediment surfaces also, Johnson appealed to lateral planation by streams. But this must be specifically excluded, for transverse stream beds, alluvia, and allied phenomena are either absent or of minor importance across most pediments. It may be that Johnson, and also Blackwelder (1931), who wrote "Pediments are graded plains due, in the writer's opinion, to the sideways cutting of desert torrents . . . Pediments are essentially compound graded flood-plains excavated by ephemeral streams", were considering special cases, but the broad sweep of pediments outward from the steeper hillslopes toward the axes of valleys is a sufficient assurance that streams do not initiate the pediment landform, nor are they responsible for its continued growth and surface form.

Howard (1942) also supported lateral stream corrasion, stressing the work done by even small tributaries which in arid regions are braided rather than channelled.

Davis (1938, p. 1367) rejected the hypothesis of lateral planation by streams in favor of planation by sheet flood and stream flood, of which he adduced many clear examples from the Basin and Range province.

Berkey and Morris (1927) were of the opinion that the pediments of Mongolia were cut by a network of shallow rills. Rill work is apparent upon virtually all pediments, but as a primary agency of pediment origin and development it is open to the objection that its continued action cuts small but definite channels which become more and more stabilized as the work proceeds. Rills and ephemeral streams are not planing agents *per se*, and pediments are normally developed as planed surfaces at right angles to the courses of the rills from the hillside to the major stream bed. Even admitting a network of rills upon the surface, their action results in a maze of small incisions into the pediment surface, not a broad planing action.

Rich (1935) has conceived of sheet wash, unconcentrated into definite flowlines, playing a leading part in the erosion of pediments:

> ". . . that where rock waste is plentiful a surface cannot be lowered below a gradient sufficient to permit the transportation of that waste; that where waste is supplied to an area from outside that area, although it may only be a thin veneer, if it is constantly removed it serves as an effective blanket to protect the underlying rocks from erosion below a gradient sufficient to permit the transportation of the debris across it; and that, other things being equal, the necessary gradient depends upon the size of the particles to be transported".

By regarding both bedrock and alluvia as weathering in their turn, the products being distributed by sheet wash, Rich believes that the form of pediments is largely governed by distribution of waste in transit under the influence of sheet wash. Rich has applied the old Davisian concept of hillslope grade to the

agency of sheet wash which, as McGee and Davis have shown, is sometimes present upon pediments. The conception is a valuable one, uniting old ideas and new, and we shall not lose sight of it. But it becomes effective only after the pediment is already in existence and does not explain the origin of pediments, nor the retreat of their commanding hillsides.

Kirk Bryan (1922; 1935; 1936), who wrote much, and authoritatively, upon pediments held that three processes combine to form pediments: (1) lateral planation by streams, (2) rill work, (3) retreat of mountain slopes. In general, also, he lays great emphasis upon the work of ephemeral streamlets in shaping pediments, especially at the inner edges. Thus: "The principal agent in the formation of pediments is the ephemeral stream, which works by incision and lateral planation. However, rainwash, rills and weathering are particularly effective in the later stages of the process".

Fair (1947), in South Africa, opined that, after the retreat of the steeper hillside under weathering, the pediment was moulded under sheet wash.

Present Hypothesis of Hillslope Evolution

The hypothesis here advanced accepts the idea that pediments originate by the retreat of scarps and steeper hillsides. No matter how, or by what agencies, a pediment is later modified; its very existence depends in the first instance upon the process of hillslope or scarp retreat. Thus any hypothesis of pediment formation must explain not only the form and remodelling of the pediment but also the reasons for parallel retreat of hillsides and scarps.

The processes of weathering and the action of gravity (*e.g.*, in forming debris slopes) have already been sufficiently elaborated and are accepted throughout the sequel; but the basis of our argument rests in the manner of water flow, derived from rainfall, over a landscape, and from personal observations and examinations made in the field at times of heavy, and of light, precipitation. The rainfalls studied vary from 3 to 4 inches in an hour to slight drizzles, and the manner of runoff varied from sheet floods down to nothing. The heaviest rainfalls I have personally experienced were 22 inches in 3 days (Zululand, 1940), 13 inches in one day (Durban, 1953), and 11 inches in one night (Knysna district, Cape Province, 1951). The results in all instances were well worth seeing; but probably the most astonishing transformation I have witnessed was in the Kalahari (December 1949) when, for 4 days continuously, the heavens opened, and the contents fell out. At the end, broad lakes stretched across that flat country almost unbroken for miles, and the Great Kalahari Desert had truly become the Great Kalahari Sea. How much rain fell altogether I do not know. The heaviest measured rainfall I have heard of in Southern Africa was in the Zomba district, Nyasaland (1946)—28 inches in 24 hours.

When rainfall is to be discharged from a landscape, the number of variable factors is large: amount and incidence of precipitation; the nature of the soil, porous or clayey, dry or saturated; detailed nature of the surface, smooth or rough; vegetational cover, sparse or close, grass or scrub; terrain, flat or steep; and so forth. These decide the relative proportions of runoff and soakage. The combinations of the various factors are infinite, and the variety of natural conditions is greater than the pen can here describe, but the several standard cases that will be narrated have all been observed over and over again in the African veld and are evidently valid for that type of terrain in which the incidence of rainfall during thunderstorms is high, where the hillslopes display clearly the several elements already discussed, and where the vegetational cover is sparse, leaving exposed notable areas of bare soil.

In what we may describe as "thunderstorms of moderate intensity", the following conditions have been observed on a normal hillside (Fig. 4). As the rain falls upon the upper parts of the slope, runoff slowly increases in volume as it moves downhill. But the volume is small, drainage is often poorly defined, and the amount of work done both by transport of soil and erosional action is not large. In fact, it would appear to be inferior to soil creep, as is indicated by the convex profile of the upper part of the slope. But, as the volume of water increases with distance from the crest of the slope and its

speed downhill increases with the steepening declivity, there comes a stage where modification of the surface under the action of running water exceeds the modification due to soil creep. This is the end of the waxing slope.

Under suitable conditions, a free face, washed clean in part by the water, now appears, and, lower down, the debris slope in turn succeeds. Both of these two elements are modified together in the hillslope development. Together they make the steepest part of the hillside, and here the ever-increasing volume of water courses rapidly down in an infinitude of tiny rills, each of which is highly turbulent and is an active eroding agent cutting back into the face of the scarp (Pl. 3, fig. 2). Such rills are normally closely spaced, the "available relief" between them is small, and hence the scarp is eroded back as a whole parallel to itself, rather like a receding wall (Pl. 1, fig. 1). Of course, not all the rills are equally large, and minor irregularities and indentations appear at intervals along the hillside. Some of these may even grow into gullies or dongas, but such irregularities tend to become partly filled with debris, and their further development is hindered. A specialized instance of this process has been termed "gully gravure" by Kirk Bryan.

The rate at which scarps retreat forms one of our later canons (King, 1944, p. 279; 1947, p. xiv).

Retreat of the steeper elements of the hillside leaves at the foot an apron of flat land extending commonly to a stream channel (Pl. 1, fig. 2). This apron of gentle declivity is the primitive pediment.

Storm water coursing rapidly down the steeper face is checked when it reaches the gentle declivity at the bottom. As the volume of runoff still mounts with increasing distance from the crestline of the hill, a much-augmented volume of water has now to be discharged across only a gentle slope. This task is accomplished (King, 1949b) partly by a change in the manner of water flow. The rills spread laterally and join up until they make a thin sheet of water of indefinite lateral extent spreading across the apron in advance of the debris slope. This change in water flow corresponds with the break in profile at the head of the pediment which is so characteristic in pedimented landscape wherever they occur. In areas of strong rocks, as we have noted, the break in profile is abrupt (Pl. 2, fig. 1); in areas of weak rocks, or where detritus and soil are thick, the break is zonal, taking place over a smooth curve. The change from hillside to pediment is sharper also in regions of high incidence of rainfall, where considerable volumes of water have to be rapidly discharged.

The pediment, a smooth landform permitting discharge in sheets over its whole area, is indeed the natural answer to the need for rapid dispersal of storm water: it is the ideal landform to dispose of the maximum volume of water in minimum time, with least erosional damage to the landscape. Pediments, once formed, may thus be expected to be relatively stable landscape features.

The manner of water flow over pediments is of prime importance. The sheets of water, at first only a fraction of an inch deep, flowing over the upper pediment gave rise in the earlier stages of my investigations to much mystification, for, though the rills coming down the steeper slopes were turbid, the thin sheets of water to which they gave rise on amalgamation were often clear, the debris having been dropped temporarily at the foot of the debris slope. The explanation of this curious phenomenon lies in the manner of water flow, for water in thin sheets often moves by nonturbulent, laminar flow. The molecules, instead of pursuing individual, interlacing, looped, and turbulent paths as they do in the thread flow of rills upon the steep hillside, join into laminae, and the laminae slide upon one another in the manner of a pack of cards thrust forward upon a table[3]. The lowest card stays in position while each card above slides progressively farther. In laminar flow, the lowest layer, in contact with the ground, has zero velocity, and the flow as a whole is nonturbulent and nonerosive. So arises the zone of clear water flowing as a thin sheet that may, under favorable circumstances, be observed toward the upper edge of the pediment.

Apparently, a special combination of circumstances is necessary for true laminar flow.

[3] See any textbook on hydraulics, under the title Reynold's Number.

The bottom must be smooth, and either of bare rock or fine debris; the sheet of water must be thin; and there must be no restriction upon the lateral spread of the water. Any influence tending to channelize the flow destroys the laminae. Under natural conditions, there is often in thin sheets a mixed flow, partly laminar, partly linear.

Across the pediment, where the volume of water to be discharged continues to increase, the sheet of flowing water becomes deeper. Commonly it also becomes turbid and erosive again. Reynold's Number once more affords the explanation for, when the sheet of water passes a certain critical thickness defined chiefly by the viscosity of water and the smoothness or irregularity of the ground, the sheets of molecules break up, and linear flow takes place again. This *sheet-flood* (as distinct from the laminar flow in sheets) is the agent that erodes and models the pediment into its typical hydraulic cross section. In addition to transporting weathered debris, it is an active eroding agent on its own account.

On the pediment, local circumstances both natural and artificial sometimes cause a concentration of flow lines in the flood, and this produces a channelling of the pediment surface. Such channelling often appears well out from the debris slope, but the gullies so produced often erode headward until they cross even the whole width of the pediment (Fig. 4). These are the typical dongas of badly eroded farmlands. Between the gullies, the pediment surface remains straight in transverse section (King and Fair, 1944).

As the storm passes and the sheet flood subsides, the pediment is covered with an enmeshed pattern of dwindling rivulets, most of them aggrading and leaving a thin surface veneer of temporary accumulation; mud, sand, or grit.

Conditions observed did not always correspond to the scheme outlined above for "storms of moderate intensity". In storms of great intensity (greater than 2 inches per hour perhaps as a generalized estimate), the zone of clear water disappeared from the upper pediment. This was presumably because the volume and discharge speed of water was so great that the rills of the steeper hillside passed over immediately into a thick sheet (sheet flood) upon the pediment, and laminar flow was thereby inhibited. At such times there was no tendency to deposit fine waste at the base of the hillside or top of the pediment. Extreme examples are sheet floods of the type described by McGee (1897, p. 100), but these are exceptional even in a desert environment and reach their maximum torrentiality only in rocky deserts of marked relief. Even in the heavy rains I have witnessed, sheet floods of the depth and violence of those described by McGee were not developed, possibly because the terrains were too absorbent. A suitable physical environment seems to be necessary, as well as concentrated precipitation, to produce extreme sheet floods. Davis has noted (1938, p. 1345) that "the time interval between successive floods on the same area must frequently be extremely long".

In times of less intense precipitation, the rills of the steep slope are prolonged across the upper pediment without joining laterally into a sheet, and tend then to erode gutters across the upper zone of the pediment. In some regions where light rains are frequently interspersed with thunderstorms the upper parts of pediments were consistently irregular and guttered in this way. Parts of Texas such as the Pecos Valley, which receives rains from the Gulf Coast as well as from the interior, may perhaps be cited, though my observations there are too limited to be certain of the example.

Clearly these observations of water flow across African landscapes need to be repeated in many other lands. Such as they are (and they were not comfortably acquired amid sheets of wind-driven rain and vivid strokes of lightning), they offer a sound basis for the understanding of hillslope elements and evolution, agreeing incidentally with the work of Wood. They show why the free face and debris slope retreat rapidly parallel to themselves, why when scarp retreat is rapid the waxing slope is poorly developed, how the pediment is modified under sheet flood, why its upper zone is sometimes irregular or guttered, and in detail how a major pediment may consist of many small pediments graded to shallow channels crossing the major, compound pediment. It explains the paradox noted by King and Fair (1944) of dongas (gullies) occurring most frequently not upon the steeper

elements of hillslopes but upon the relatively flat pediments as due to breakdown of the laminar flow conditions into linear flow with the inevitable production of channels. Lastly, it shows why the long concavo-convex slopes of certain northern lands must be regarded as degenerate, evolving no longer by scarp retreat but more or less atrophied. Their active elements, the free face and debris slope, are lost; the waxing slope produced by weathering and soil creep forms apparently at an inordinately slow rate compared with scarp retreat; and the active pedimentation which perhaps formed the original concave slope is now arrested so that this part of the slope now passes downward by gradual transgression through progressively less weathered detritus into solid bedrock. These features are well shown in chalk areas where runoff is minimized by percolation underground.

But more data are necessary before we can carry the conception much further. Before leaving the subject, however, we may add one hint useful for field observations: To distinguish laminar flow from linear flow, it is only necessary to observe the manner in which the water passes an obstacle such as a pebble or the stem of a small bush. If a depressed cone appears in the water surface about the obstacle, the flow is laminar and nonturbulent, if on the contrary the water surface piles up against the face of the obstacle the flow is linear and turbulent.

Convexity of Upper Slopes

As we have seen, extensive and widespread convexity of upper slopes is a sign of degeneration in the hillslope cycle; it means that soil-creep, slow and smothering, is dominant with weathering over the normally more active process of rain- rill- or sheetwash. This tends to be so more in areas of low than of high relief where the waxing and waning slopes can more easily meet, with elimination of the more actively disposed free face and debris slope. It appears also under climates where rainfall is moderate and evenly distributed, and runoff is consequently at a minimum; and in regions where a turf or sod is developed by the binding action of roots. Lastly, it appears more generally in areas of weak than of strong rocks. In short, bicurved slopes appear in situations where deep weathering is promoted and where the products of weathering are removed but slowly.

Fair (1947), Davis, and Rich, arguing from different premises, have concluded that at the final stage of an erosion cycle the ridge crests should generally become convex in cross section. But we have to record, *e.g.*, from Southern Rhodesia, that convexity, even at the ultimate stage of the cycle is not universal. In the hard, granitic terrains of that land, biconcavity sometimes persists upon pedimented ridges until the opposed pediments meet at a noticeably angular crest. Small castle-koppies (tors) commonly surmount such ridges, and these still maintain angular relations with the pediments at their bases (Pl. 3, fig. 1; Pl .4, fig. 1). Again, pronounced convexity appears to be a sign of degeneration in the hillslope cycle.

As shown in a later section, pronounced convexity of hillslopes under soil creep is apparently a recent geological phenomenon. Prior to the mid-Tertiary, it exercised but a minor role in landscape evolution. Four general explanations of this new emphasis have been offered:

(1) The usual textbook explanation is that these pronounced convexities are related to a more humid climate. Insofar as more humid climate promotes deeper weathering and a greater thickness of soil mantle, this would allow of more soil creep not only on hilltops but on the lower country as well. *Per contra* the superb hydraulic curve of pediments—the acme of hillslope form under running water—is characteristic of semiarid and arid environments, so that there is a *non sequitur* somewhere. Even consideration of the incidence of rainfall, evenly distributed or episodic, does not remove all the difficulties, and we must conclude that the hypothesis in its present form is inadequate.

(2) Modern French authors have referred summit convexities to a periglacial regime. Pronounced concavo-convex slopes do in truth show a distribution about the areas of Pleistocene glaciation, but this does not of itself explain the convexity. Professor Hollingworth of University College, London, has privately made the suggestion that the deeper weathering indicated by convexity is connected with deeper

frost penetration. It is to be hoped that he will follow this suggestion up.

(3) Late Tertiary time has been punctuated by epeirogenic rise in all continents, with spasmodic uplifts following one another apparently at closer and closer intervals. River incision may thus have been accelerated from stage to stage with the production of steeper gradients on lower (*i.e.*, convex) slopes.

If this were so, then steepening should continue to the bases of hillslopes near the mouths of rivers. Other objections to the hypothesis can also be raised, but it may serve in combination with other factors.

(4) The fourth hypothesis stems from the observation that convexity of hilltops becomes notable from the mid-Tertiary onward, and links this with the great radiation of the carpet- or turf-making grasses during and after the Miocene period. With the production of turf or sod an entirely new factor entered the development of landscape. Surface runoff was hindered, and the development of the water-cut profile was minimized; percolation was enhanced, and deeper weathering and soil creep resulted. So the parallel retreat of slopes with the production of pediments was reduced while the waxing slope was enhanced. In areas of slight relief, or of weak rocks, the free face and debris slope disappeared, smooth concavo-convex slopes became almost universal; and so there came about a fundamental change in the aspect and perhaps the manner of evolution of the landscape.

Homologies between the Landscapes of Humid Temperate, Semiarid, and Arid Regions

As soil removal by wash and creep is still greatest on the steepest part of slopes, namely where the convex and concave slopes meet, it is likely that scarp retreat is not wholly prevented where a sod mantle is present, but the rate at which it occurs must be very much reduced from that upon bare rock and talus.

The manner of scarp retreat, if it exists, in turfed regions is a subject for study. The concave lower hillslopes in such regions do not differ essentially in profile from many pediments, and some which I have examined both in Britain and in northeastern North America do not exhibit a continuous section downward by gradation into bedrock but were apparently at one time cut-rock surfaces, though they seem to have atrophied so that weathering now affects the originally smooth, planed rock surface. Such changes are not inconsistent with the lying of snow during the Ice Age, instead of the continued dominance of running water.

Evidence of associated scarp retreat to form these probable pediments must also be sought. Fair (1948), studying slope form and development in the humid coastal belt of Natal, inquired whether smooth scarps in the Table Mountain Sandstone country, where the free face had been transformed into a waxing slope thinly covered with soil, still retreated in the ordinary way or whether they lost declivity until they were finally obliterated. He concluded that, apart from sections of the scarps which were gashed at intervals by gullies,

> "observation suggests that the compound slope, convex above and concave below, assumes a gradient upon which the rate of supply of weathered waste is balanced by its rate of removal by gravity and sheet-wash. Once scarps are rounded off to this gradient, no further loss of declivity occurs as long as the unconsumed (sandstone) remains. It will be seen that the resistance to weathering of the sandstone is partly responsible for this and that slopes on other, less resistant rocks, might continue to flatten with time".

In Britain it has long been customary to speak of the retreat of the Chalk scarp (*e.g.*, west of Salisbury). Much of the British countryside is, indeed, notoriously scarped, and data concerning the lower slopes would be welcome.

In the Appalachian Mountains of North America, transversely concave valley bottoms often occur and not flat flood plains. Do these valley bottoms have concave cut-rock floors? Davis (1930) has quoted valley floors with rock basements "in abundance along the belts of less resistant strata in the folded Appalachians from Pennsylvania and Virginia to Tennessee and Alabama". These seem to differ from pediments only in that nobody has yet called them pediments. This applies also to details of the earlier Appalachian surfaces, as Davis goes on to show[4].

[4] *Fair comment!* "See my earlier note that Davis saw pediplains but called them peneplains".

It seemed to me also that certain of the ridge flanks did not coincide with the structure nearly so well as is required by the usual conception of the evolution of Appalachian topography by downwearing under weathering and stream action. Several instances were noted west of Harrisburg where slope angles did not agree with dips of strata, or where the waning slope had transgressed appreciably across the geological boundary from the soft onto the hard strata.

There is likewise the retreat of the Catskill Mountain front from the Hudson Valley which does not seem wholly explained upon structural grounds but requires active scarp retreat. Instances might be multiplied, but my own observations have been too fleeting to command respect, and the research must await the attention of some investigator more leisured in the region.

In the same outstanding paper Davis (1930, p. 136) drew attention generally to the presence of pediments or pedimentlike features in landscapes developed under humid as well as arid climates, thus demonstrating their essential universality in landscapes of subaerial origin. This was an advance toward harmonizing what had been thought of previously as two distinct modes of landscape evolution, and his descriptions of these features and their mode of origin are so important that we quote them *in extenso*. Writing of the valley floors of humid regions, Davis noted that as the valley-side slopes

> "... recede from the banks of the graded stream to which they previously descended ... narrow strips of valley floor ... will be developed at their base, back of whatever flood plain ... is simultaneously formed by the stream. As these strips widen, the concavity of the profile across the valley bottom ... is given broader expression" (Figs. 6, 7).
>
> "The lateral valley-floor strips are, like the flood plain, underlain with degraded rock which may be called the valley-floor basement. It has a somewhat ragged and vague surface, except that where cut by lateral shifts of the stream it may be more even and better defined. The lateral strips will be everywhere covered by detritus, partly derived from local subsoil weathering, partly washed down from the valley-sides; and the detrital cover will, in time, constitute a large part of the valley floor.... Each lateral strip of the valley-floor will have, besides its downstream slope, a somewhat stronger transverse slope from the valley side towards the stream or its flood plain....
>
> "The lateral swinging of the graded stream broadens the flood plain and its rock basement ... and thus the valley-floor strips may be encroached upon; but in spite of this *they gain width by the retreat of the valley-side slopes*" (Italics mine).

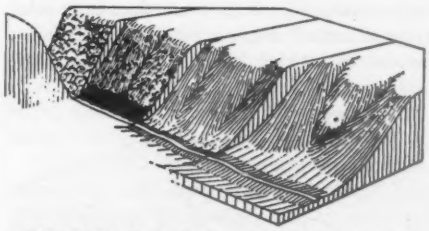

FIGURE 6.—DEVELOPMENT OF ROCK FLOORS (PEDIMENTS) UNDER A HUMID CLIMATE
According to Davis (1930)

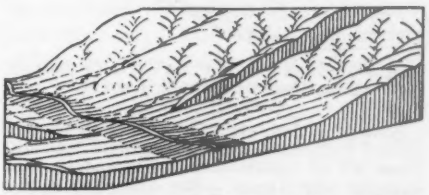

FIGURE 7.—CONTINUED DEVELOPMENT OF ROCK FLOORS (PEDIMENTS) AND INCIPIENT SECONDARY PEDIMENTS UNDER HUMID CLIMATE
According to Davis (1930). Note in Figures 6 and 7 the semiparallel retreat assigned to the hillslopes.

Davis also noted that

> "In the meanwhile many brooks, short side branches of the main stream, will excavate ravines in the valley sides (Fig. 7).
>
> "After the side-brook ravines are developed to the stage of having their own flood plains,[8] side branches of the main valley-floor strips, each with its underlying rock basement, will be opened along their margins heading farther and farther up the ravines".

Under humid regimes, Davis has here completely admitted scarp retreat with concomitant pedimentation—for what are these rock-floor strips but pediments? (Compare for instance the pedimented landscape of Pl. 2, fig. 2 with Davis' descriptions.) Davis seems to have been prepared to accept them as such, and from long experience in the field I have come quite independently to the same conclusion. They are as homologous with pediments as monadnocks are with inselbergs. Thus I can see no reason for denying to the concave rock floors of humid regions the status and name of

[8] I would say even before the flood plains develop.

pediments. Their development may be less extensive and more imperfect, and they may be more obscured beneath debris than their relatives of semiarid regions; but of their origin by retreat of backing hillsides, and subsequent modification under running water, both sheet and rill flow, I have no doubt.

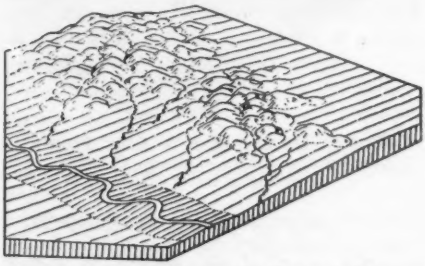

FIGURE 8.—LATE STAGE OF THE CYCLE OF EROSION UNDER HUMID CONTROLS ACCORDING TO W. M. DAVIS

Note particularly the wide rock floors, concave upward (pediments), and the steep-sided residuals. These are all typical of pediplain and contrast markedly with Davis' earlier conception of the peneplain with its supposed broad convex divides and faint residuals due to structure only.

Fair (1948a, p. 44) has contrasted the pediments of the semiarid interior of Natal with those of the more humid coastal belt. His conclusions may require some modification in the light of the hypothesis of states of water flow enunciated herein, but remain a worthy reference for students of the subject.

Kirk Bryan, too (1940, p. 266), considered hillslope development in arid, tropical, and humid temperate climates and came to the preliminary viewpoint

> "(a) that slopes of hills are characteristic of the climate and the rock; (b) that these slopes once formed persist in their inclination as they retreat; (c) that they disappear only when all the volume of rock above the encroaching foot-slopes or pediments has been consumed".

In his 1930 analysis, Davis has gone a long way from his simple belief in "graded" hillslopes of the Geographical Essays. His exposition ends indeed with a diagram (Fig. 8) of the ultimate landscape form under the erosion cycle which is far removed from his earlier concept of the classical peneplain. It is none other than the pediplain, typical senile landscape familiar in semiarid environments, utterly unlike the peneplain of broad convexities in that it is composed of numerous broad, concave pediment or pedimentlike surfaces from which rise steep-sided residuals, not gentle monadnocks. All this accrued from his two-fold recognition of (1) rock basements or pediments and (2) scarp retreat in humid temperate regions.

TABLE 2.—CHARACTERISTICS OF MATURE PENEPLAIN AND PEDIPLAIN

Peneplain (Fig. 9)	*Pediplain* (Fig. 8)
Broad flood plains of rivers	Relatively narrow flood plains of rivers
Divides invariably subdued, and broadly convex	Divides concave on both sides or only narrowly convex
Residuals gentle, convex	Residuals may be abrupt, concave in section
Concave profiles only in lower parts	Concave profiles of broad sweeping pediments dominant in the landscape
Origin by slope flattening	Origin by scarp retreat and pedimentation

Yet, with the main principle of the fundamental identity of all epigene landscapes firmly within his grasp, and with the inference that the semiarid landform with widespread pediments is the norm of epigene landscapes clearly before him, Davis relapsed into the old ideas of progressively flattening hillslopes (1932) and the consequent *idee fixe* of the peneplain. In doing so, we believe, he turned from the gateway to interpretation of the various regional cyclic histories of the earth (King, 1950).

PENEPLAIN VERSUS PEDIPLAIN

The reader is by now familiar with these two similar, but fundamentally distinct conceptions of the ultimate or senile landscape resulting from epigene erosion (Fig. 1; Table 2).

The distinction between the two types of landscape shows best in tough granitic terrains and is least marked in regions of weak shales and limestones.

Let there be no compounding upon this point: Though both landscapes may conform to the loose definition of "a landscape of low relief re-

sulting from prolonged subaerial erosion" they are not, nor at any stage ever were, the same thing.

If the prime agents in each case, slope flattening or scarp retreat and pedimentation,

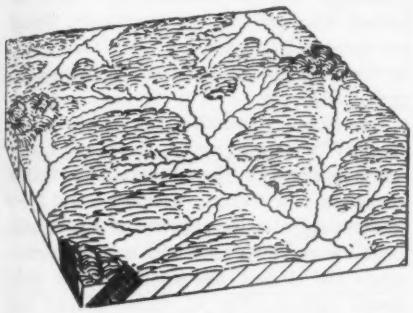

FIGURE 9.—STANDARD CONCEPTION OF THE PENEPLAIN

Showing flood plains of rivers, broadly convex divides, and exact correspondence of monadnocks with boundaries of resistant rock formations (After Cotton).

operate almost exclusively, then one form is valid and the other is not. If both are regionally operative as, for instance, slope flattening in humid-temperate regions and parallel scarp retreat in semiarid regions, then peneplains may be expected in the former and pediplains in the latter environment.[6] What has Nature to say on this? Do we find both forms, or either exclusively?

Before answering this question we shall consider an opinion of Walther Penck whose concept of senile landforms agrees with the pediplain. This form he called an *Endrumpf*. But Penck recognized also landscapes in which convex slopes appeared owing to gentle uplift and weak incision of the rivers. This landscape type he deemed to be initial in a new phase of erosion and termed it a *Primärrumpf*. He related both landforms to land movement either slower or more rapid than the rate of erosion; I have preferred to think of them in terms of the erosion cycle as a pediplain and a redissected pediplain (two-cycle) respectively—that is, as landforms pure and simple, the former is an end-product of the erosion cycle, the latter, after rejuvenation of the rivers, is the initial stage in a new, second cycle of erosion.

Davis (1932, p. 419) discussed Penck's concept of *Endrumpf* and *Primärrumpf*, but was handicapped in his judgment by his long-held belief in a convexity of old-age landscapes.

My own observations upon old-age surfaces permit very clear and definite conclusions in my own mind. All the truly old-age surfaces I have seen, and I have studied and mapped quite a few, were multi-concave in form, and bore residuals that were steep-sided and concave in profile themselves. The topographic boundaries of such residuals often do not coincide with geological boundaries between resistant and weak rocks. Even if the residuals are of resistant rocks, the topographic discontinuity between the residual and the adjacent plain had almost invariably transgressed from the weak onto the resistant formation. The landforms were pediplains or *Endrumpfe*.

Convexity occasionally appeared along the crests of some of the divides especially in areas of weak rocks such as Karroo shales, but it is clearly a feature induced only as senility advances, and was subordinate to concavity in the landscape observed as a whole. It was always summit convexity only.

Convexity in the neighborhood of the streams upon an old-age land surface, however slight, was found to be an infallible token of incision controlled by the arrival of a new cycle of erosion (Pl. 4, fig. 2), and if the observer follows downstream the amount of incision generally increases, and the second cycle becomes obvious. Below the region of maximum incision, youthful convexity of course decreases again in the second cycle, and concavity appears at the bases of hillslopes.

Judged on these criteria, observed in nature, I must declare that I have seen pediplains and diagnosed dissected pediplains, *i.e.* features akin to Penck's *Endrumpf* and *Primärrumpf* but I have never seen a peneplain. To arrive at this conclusion, I have scrupulously re-

[6] Dr. T. J. D. Fair comments "Davis's peneplains were described mainly from humid temperate climates where, with heavy vegetation and deep soil, creep probably operates more obviously than in semi-arid climates. Thus, just as the pediplain of semi-arid regions probably displays greater summit convexities in humid-temperate regions, so probably do Davis's peneplains of humid-temperate regions display smaller summit convexities in semi-arid areas".

jected from consideration numerous planed landscapes in the history of which doubt existed (*e.g.*, of resurrection), or where the second cycle had advanced much beyond the incipient stage. I have also neglected many with essentially straight hillslopes where stream incision and hillslope development appear in a delicate state of balance. They do not seem to afford exceptions to the theory. Fair has measured several slopes of this type.

Slow uplift is not essential, in my opinion, to the generation of a *Primärrumpf* from an *Endrumpf*, but slow incision of the streams, from any cause—*e.g.*, gentle tilting toward a basin—brings out the relation.

"Normality" in Landscape

Textbooks of geomorphology commonly refer to the following types of erosion cycle: the "Normal Cycle" of Erosion typified in humid-temperate lands, the Semiarid Cycle of Erosion, and the Arid Cycle of Erosion. From this use of the word "Normal" we dissent.

The humid-temperate climates of Europe and North America have been taken as "normal" because they are typical of the home regions of western civilization, they appear abnormal to dwellers of [for instance] semiarid regions who are used to long periods of stable, clear weather and are apt to refer disparagingly to the fickle weather of northern climates. Those in semiarid regions might legitimately point out also that, whereas the action of running water can be plainly seen and studied in landscapes of the semiarid environment, with its areas of bare ground between tufts of grass, the processes of surface wash are impeded in the northern lands by either a close carpet of grass or by woods which, until recently, were even more widespread. Deforestation has become universal in both Europe and North America only since the Middle Ages. Moreover, the northern parts of both these areas were earlier covered by the greatly extended glaciers of the Pleistocene. Even those areas which were not covered by ice sheets suffered a periglacial regime of permafrost and snowfall instead of rain. A superabundance of waste upon the hillslopes and in the valleys alike was but one of the consequences of this regime.

We ask ourselves therefore, are these regions to be cited as "normal" and serve as a standard of comparison for geomorphic processes and evolution?[7] On the contrary, thoroughly abnormal standards were gratuitiously, though not unnaturally, assumed in the classic studies of the subject, and have been perpetuated uncritically since.

An unbiased approach to the use of the word "normal" is now necessary. Davis and Cotton (1941) regarded any departure from the humid-temperate as a "climatic accident", and Davis (1930), even after he appreciated the homologies between landforms under epigene environments and derived the pediplain even under humid regimes (Fig. 8), found himself confirmed in "... an earlier belief that the cycle of arid erosion may be, with all its peculiar consequences, reasonably treated as a variant of humid erosion". Nowadays an unbiased observer, studying the world as a whole, might very well arrive at a very different conclusion.

We seek to establish what is "normal" in epigene landforms. Let us be quite clear upon one initial point: The student who would learn the normal processes of landscape development, and the normal types of landforms, and the normal manner of scenic evolution *must study them in semiarid regions*. There will he see, in regions of exoreic drainage, the type of rainfall and stable climate which characterizes most of the land area of the globe, and the manner in which this rainfall is dispersed. There generally will he avoid serious complications introduced by the climatic fluctuations of the Pleistocene. There will he see best displayed the elements of hillslope form, the most active development of landscape under scarp retreat, and the processes of pedimentation, in short the most actively evolving type of landscape (Fig. 10).

[7] Since this was written I have seen the scheme of Morphogenetic Regions based upon climatic type proposed by L. C. Peltier (1950). He lists nine regional types each "normal within its own regime and a 'climatic accident' only when it temporarily encroaches upon the area of another regime". This novel and fundamental point of normality is more far-reaching than that of the text which might then perhaps be termed "mean or average cycle".

A deviation from the semiarid mean either toward greater humidity or greater aridity provokes the accumulation of waste in the landscape, symptomatic of a decreased efficiency in landscape evolution. On the one hand is

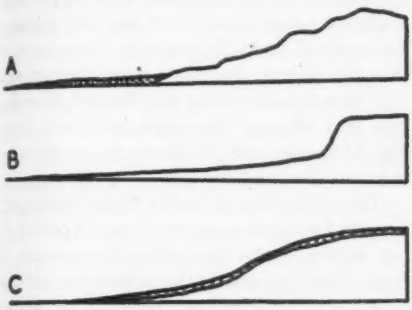

FIGURE 10.—LANDSCAPE PROFILES UNDER ARID, SEMIARID, AND HUMID INFLUENCES

(A) Broad benches and structure lands with bahadas, (B) Scarps, pediments, and inselbergs, (C) Smooth concavo-convex slopes leading to valley plain and rounded hill outlines.

produced a soil blanket that mantles perhaps even the free face and debris slope, with dire results upon the active evolution of the landscape as a whole; on the other abundant waste accumulates in the hollows as fans and bahadas, so preventing the operation of the usual processes of pedimentation which are responsible for the finishing touches put to a landscape. In extreme cases, even the exoreic drainage system may be destroyed, and convex rock surfaces form as base to the alluvia instead of concave pediments. (*Cf.* Lawson, 1915.)

So much for modern landscapes; what was "normal" in the geologic past? All the Mesozoic and early Tertiary erosional landscapes of which I have accurate knowledge were pedimented "semiarid" landscapes, even though they occur in what are now regarded as "humid" regions. Not only regions such as Africa, the Gobi, and Australia were pedimented at those dates but also Europe and North America. Davis has mentioned, for instance, the concave rock floors of the Blue Mountain region of the Appalachians, while Baulig's descriptions of the Oligocene landscape of the Central Plateau of France leave us in no doubt regarding the pedimented nature of that former landscape. Even the small and difficult example of Dartmoor, Southern England, has the characteristic curve of pediments when the oldest ridge crests are viewed in profile, and the famous tors are identical with the castlekoppie residuals of the Rhodesian veld. There, apparent to the trained eye, was anciently a landscape which had reached the penultimate stage in the pediplanation cycle. The dissection of the ancient summit pediplain is far advanced, and indeed a second partial cycle was already developed presumably before the end of the Tertiary. This younger landscape, still older than the main dissection of the region, stands 300–500 feet below the original surface around Postbridge, and itself shows small local pediments.

Only since the middle Tertiary, apparently, have landscapes of "humid" type become prominent, probably as a result of the radiation of carpet grasses. We shall not err if we visualize the surroundings of the Mesozoic dinosaurs and the early Tertiary mammals as landscapes of semiarid type, which were widespread and normal in those times.

We have been considering continental landscapes of erosional type; what evidence is afforded by continental landscapes of depositional type? Therein the same standards regarding climatic normal of the geological past are manifest. Few continental sequences other than those of the Ice Age show evidence of deposition under the "humid" regime. The great majority on the other hand offer abundant indications of deposition under prevailing "semiarid" to "arid" conditions. The Karroo system with the mighty Beaufort Series, reptile-bearing and with correlatives in several continents, or the Devonian and Triassic red beds of the northern hemisphere and the Tertiary sequences of the Midwest of North America are all typical of accumulations in the semiarid areas of to-day.

In seeking fundamental geomorphologic standards, it is therefore necessary to reject almost all existing conceptions about "normal processes" and "normal cycle of erosion". Thoroughly abnormal standards involving the arthritic concavo-convex hillslopes, crippled and reduced to impotence beneath a blanket of

sod, were gratuitously assumed half a century ago. These conceptions of normality require revision. It may be difficult, and even painful, for geologists resident in the "humid-temperate" regions to revise their opinions on this topic when all their textbooks tell them otherwise and the landscapes about them appear to confirm this judgment. But over the world as a whole, during the geologic past as well as the present, the facts are inexorable, and the "semi-arid" type of landscape must be regarded as the normal, most completely developed, most actively evolving, standard type of landscape.

Epigene Cycle of Erosion

The secret of landscape evolution lies, evidently, in the mode of development of hillslopes. Now hillslopes may be initiated either as valley sides consequent upon stream incision, or as tectonic features the result of faulting, monoclinal warping, or even gentle tilting in a landscape. This is an important distinction upon which the evolution of the landscape as a whole afterwards depends.

In the first case, after a region has been uniformly uplifted, nickpoints run rapidly up the rivers and tributaries so that stream incision and the production of youthful valleys appear powerfully in the landscape. The interfluves are at first scarcely affected, and, if the initial surface was a plain, there comes into existence a landscape of the dissected plateau type, such as the Appalachian plateaus around Pittsburgh, Pennsylvania. Where the streams were already organized in a previous cycle upon structural controls, an incised trellised pattern develops like that of the folded Appalachians, or the joint-controlled pattern of the Matopo granite area, Southern Rhodesia.

The second case is illustrated by Southern Africa where continental uplifts terminated near the coasts in great monoclines facing toward the sea. These tectonic scarps are directed athwart the drainage lines (*e.g.*, Natal, Fig. 3). Once generated, the scarps that are large and therefore retreat rapidly make a series of risers which separate different cyclic land surfaces ascending step-wise successively away from the coast. The scarps retreat virtually as fast as nick-points advance up the rivers, so that the distribution of successive erosion cycles bears no relation to the drainage pattern whatsoever (Fig. 3). Under erosion, the scarps may not always remain clean and wall-like. Some degenerate into zones of dissected country separating an upper and earlier planation from a lower and later surface; but many of the cyclic scarps of Africa impress precisely because they have not so degenerated but have remained clear and distinct throughout their history. The oldest date even from the Mesozoic, and these are no exception—*e.g.*, the Drakensberg (Pl. 1, fig. 1).

The destruction of earlier cyclic landscapes may be less rapid under the retreat of continental erosion scarps than when the country is first divided up under stream incisions and the proportion of hillslopes available for active retreat is thereby greatly multiplied. In this factor probably lies the explanation for the great ages of many African landscapes as compared for instance with the cyclic landscapes of North America. Smooth erosional landscapes such as the South African highveld or the Rhodesian plateau belong to the Mesozoic cycles, yet standing above them and separated by steep scarps are remnants of even older, conceivably Paleozoic, planations, some of which have apparently been exposed continuously to the weather since that time—*e.g.*, southern Congo.

Nothing like this appears to exist in North America, where few traces of very ancient cycles remain unless their preservation has been aided by burial for a time. A Mesozoic surface is described from Minnesota, and there are one or two other probable remnants of Cretaceous landscapes, but the fundamental surface from which the eastern half of the United States was later carved is the mid-Tertiary land surface, found emerging from beneath the Ogalalla covering formation in the Llano Estacado of the West, and appearing as the summit uplands of the Appalachian Plateaus, Schooley "peneplain", and the New England upland in the east.

We may now follow certain points in the evolution of landscape after the youthful stage. When the streams are already well and widely incised, parallel retreat of hill and valley sides reduces the areas of initial surface

upon interfluves. If the relief is low the natural curve of the pediment may soon take it up to meet the waxing slope, so that concavo-convex valley sides appear and slope development becomes moribund. If, on the other hand, the relief is high, shrinkage of the interstream plateaus continues until opposed pediments meet and a biconcave transverse section to the interfluve results. A noteworthy fact emerges here. The upper surface of the interfluve remains in this case virtually without alteration until the hillslope elements of the new cycle encroach upon it. I have seen on outliers of upland surfaces quite unmistakable summit areas of as little as 2 acres surviving virtually without alteration though all the country for miles around had been consumed by scarp retreat in a newer cycle of erosion. The hill called Showe in the Shamwa district of Southern Rhodesia is a superb example of this. Where jointed granites form the terrain, and hillslopes stand steeply, bornhardts are a typical landform under this process (Pl. 3, fig. 3).

Where primary dissection of the country is not by stream incision, but by the retreat of great cyclic scarps (Fig. 3), earlier cyclic land surfaces may survive long after the initiation of the new cycle. They remain practically unaltered until destroyed utterly by the encroachment of the fresh cyclic scarp (Pl. 1, fig. 1). Here is one of the most important canons of landscape development.

Whereas the cyclic erosion of the older Davisian conception, with its emphasis upon slope flattening and universal downwearing under the action of weathering, early obliterated all trace of the initial surface and reduced everything to a peneplain, the newer concept involving cyclic scarp retreat permits the initial surface, itself a record of an earlier cycle of erosion, to remain much longer in the landscape. Inherent in the doctrine is the persistence of the older surface without significant alteration for a long period into the currency of the new cycle. Two cyclic surfaces are for long co-existent, the older above the retreating scarp and the younger below.

The far-reaching consequences of this have already been demonstrated in studies of African and other continental landscapes (King 1949a; 1951b). It forms indeed the primary canon for deciphering the past histories of landscapes, and with its aid the first, tentative correlations of dominant cyclic landscapes have been made from continent to continent. On the down-wearing hypothesis, leading to Davisian peneplains, the very possibility of this achievement would be denied.

After scarps have retreated, both the major continental cyclic scarps transverse to the drainage and the minor scarps or valley sides, a wide development of pediments becomes apparent in the landscape. From either side of every stream large and small twin pediments extend laterally and make by coalescence pediment plains upon which stand unconsumed, steep-sided residuals (Pl. 4, fig. 3). In Southern Rhodesia (excluding the sand country of the west) pediments constitute probably 65 per cent of the landscape; the rest is residual hill masses. As the pediments provide the areas for cultivation their social and economic importance is also great.

As they widen they are subject to regrading, especially the steeper part toward the foot of the overlooking scarp. From an initial slope of perhaps 10°, they are reduced commonly to declivities of half a degree or so. This regrading is accomplished chiefly under sheet wash, and even ancient and very flat pediments often show beneath a thin veneer of transported material a smoothly cut, unweathered rock surface. Fair remarked of the general smoothness of pediments in the semiarid Karroo that "sheet-wash and rills are capable of regrading them with little difficulty in response to minor changes of base-level". But such changes are not necessary to regrading as Rich (1935) has also noted: "Drainage diversions and the dissection of abandoned fans and pediments are normal and to be expected. They do not require the intervention of diastrophic or climatic changes".

Conflict of pediments from neighboring streams, in which the stronger (usually the larger) extends its area at the expense of the weaker, is also a very real process causing regrading. The "development of master pediments", as we may call it, has most important consequences when we come to the dating of land surfaces.

In the later stages of the cycle, when the hills

are reduced to small, rocky koppies and the pediments stretch perhaps for miles (Pl. 4, fig. 3), a *multi-concave* landscape is characteristic. In Southern Africa, this advanced evolutionary stage is widely developed with respect to stable, or gently falling base levels. Extensive alluvia or even flood plains are seldom present, and the soils may be quite different on opposite sides of even small streams. We may describe as typical the landscape on the east and south sides of Manda (Concession), Southern Rhodesia, where the pediments sweep right down to the swampy area of the streams themselves. The Marodzi valley west from Concession is larger and possesses a strip of flood plain 200 to 400 yards wide. But the bordering pediments are each over a mile in width before they abut against the rocky slopes of koppies. The general slopes of these koppie sides are nearly 30° inclination, but the lower 300 to 500 feet may approach a slope of 45°. The inner, middle, and outer zones of the pediment have slopes of 1°40′, 1°10′, and 30′ respectively, and these estimates are more or less standard for the district. Where the pediments pass as rock fans into the mouths of narrow side valleys they become steeper, sometimes 5° or 6°. The pediments themselves do not seem to exceed slopes of 3°, even when they are narrow and not very concave. In the Glendale-Bindura district of the same colony, the Mazoe Valley, though miles in width, nowhere possesses extensive flood plains but always a narrow river channel with wide, complex bordering pediments. Small residual hills rise from between the various pediment surfaces related to sundry tributaries. And so the descriptions of maturely pedimented landscapes in Africa could be repeated for innumerable instances.

At senility a continental erosion surface is not simple. It consists of a complex assemblage of closely related surfaces (chiefly pediments) referable to a multitude of drainage lines and basins, large and small, and united only in reference to a single, widespread, long-stable base level.

Even in this stage there may be further, accidental complications such as the multiplication of cyclic surfaces by structure (split nickpoints), local land movement, and so forth; but these factors have been discussed separately (King, 1947; 1951b) and shall not detain us here.

Observational data suggest that even at the senile stage there is little weathering of the firm rock beneath the pediment veneer, but this conclusion may not be valid. In the African theater, Pleistocene climatic changes were expressed in an alternation of arid and relatively more humid phases. As has been remarked by Cooke, the superficial materials overlying rock-cut pediments, when revealed in the sides of dongas, in many places contain Middle Stone Age implements. Following a previous phase of pedimentation there was thus

PLATE 1.—EROSIONAL SURFACES AND SCARPS IN SOUTH AFRICA

FIGURE 1.—NATAL DRAKENSBERG SOUTH OF CHAMPAGNE CASTLE

View north. Relief over 4000 feet. This scarp began at the Natal coast as a structural feature, the Natal monocline, in the late Jurassic. It has since retreated westward under erosion the full width of Natal, 150 miles. Its steepness is maintained by innumerable gully heads. The ancient, Mesozoic landscape above has remained virtually without alteration throughout the late Mesozoic and Tertiary. Anon. photo.

FIGURE 2.—TYPICAL LANDSCAPE OF A SEMIARID REGION, THE KARROO, SOUTH AFRICA

There is no flood plain, and there are wide pediments backed by steep hillslopes on which are minor structural effects caused by horizontal, resistant sandstones and dolerites. T. J. D. Fair photo.

PLATE 2.—EROSIONAL FEATURES OF GRANITE TERRAINS, SOUTHERN AFRICA

FIGURE 1.—SMALL GRANITE HILLOCK

Showing waxing slope, free face, and incipient pediment. In the absence of detritus the sharp angle between free face and pediment is clearly shown. Hluhluwe Game Reserve, Zululand.

FIGURE 2.—RETREAT OF HILLSLOPES AND GROWTH OF PEDIMENTS

In a region of jointed granite (Matopc Hills, Southern Rhodesia). Airphoto from 15,000 feet. Aircraft Operating Co. photo.

FIGURE 1

FIGURE 2

EROSIONAL SURFACES AND SCARPS IN SOUTH AFRICA

FIGURE 1

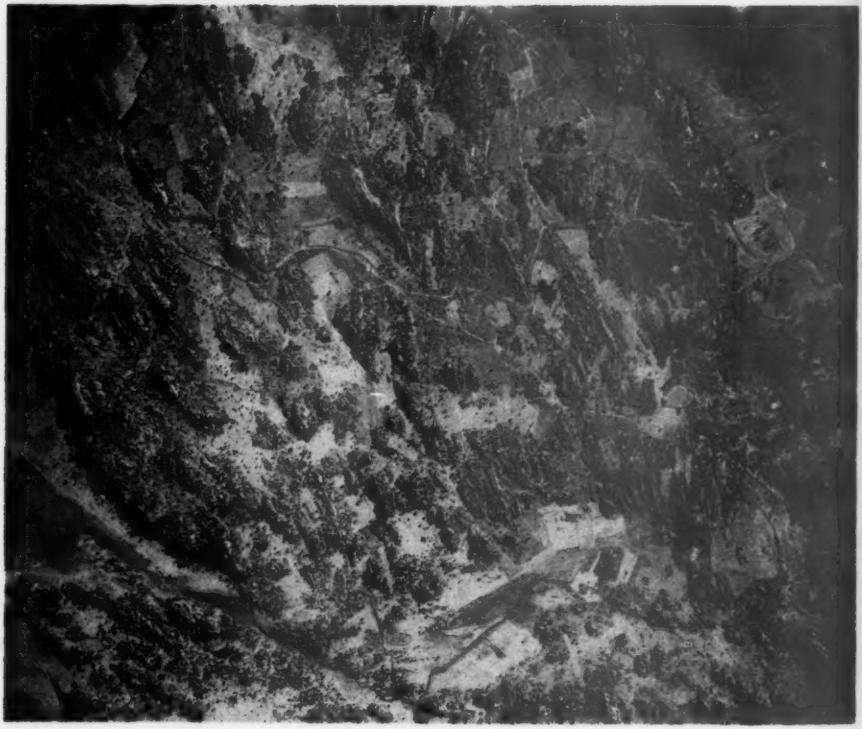

FIGURE 2

EROSIONAL FEATURES OF GRANITE TERRAINS, SOUTHERN AFRICA

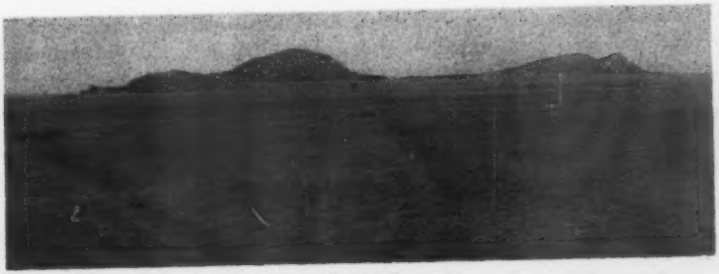

FIGURE 1

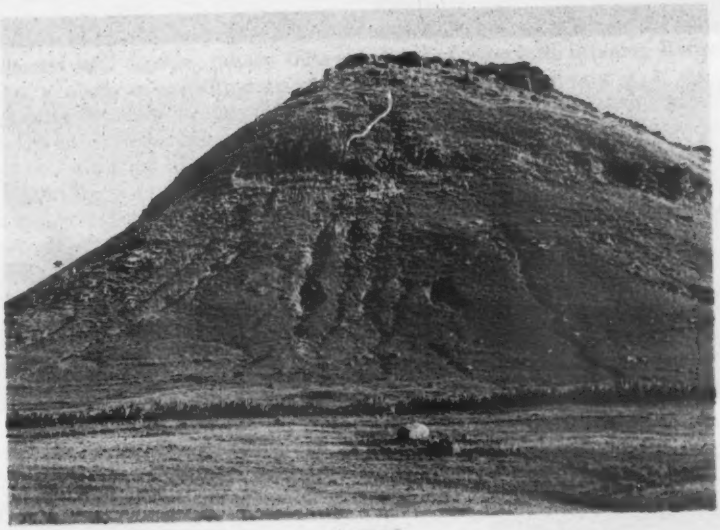

FIGURE 2

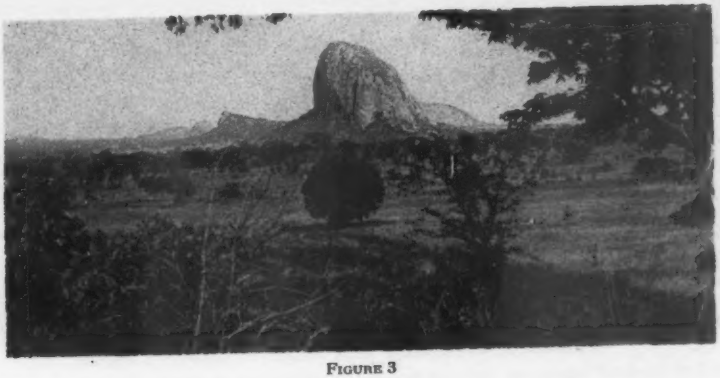

FIGURE 3

RESIDUAL HILLS IN SOUTHERN AFRICA

FIGURE 1

FIGURE 2

FIGURE 3

PEDIMENTED LANDSCAPES IN SOUTHERN RHODESIA

a widespread phase of deposition while the Middle Stone Age folk occupied the scene. Still later came the phase of incision when the dongas were trenched through the depositional blanket. All these changes, covering an enormous area, appear to have been climatically controlled. They obscure the ultimate stage of pedimentation over much of Africa.

Dating of Land Surfaces

Land surfaces are customarily dated by the discovery of superficial deposits whose age can be defined. But with the regrading and conflict of pediments such deposits, usually thin anyway, may be wholly or partially removed. The net alteration in the landscape produced by one pediment cutting shallowly across another may be small indeed and quite irregular, but from the point of view of dating it becomes highly important.

Let us assume that a pedimented landscape of low relief except for residual koppies and small plateaus acquired a thin cover, partly detrital and partly residual, of Cretaceous age. The original landscape must therefore have been either Cretaceous or pre-Cretaceous. The region is not invaded from outside by incisions or other features of later continental erosion cycles, but continues to develop solely by the processes of pedimentation under which certain struggles for mastery go on, more-favored pediments cutting shallowly across their neighbors and so removing the cover and perhaps a few inches or feet of bedrock. Over the area so replaned, Eocene deposits may accumulate. So from time to time during the Tertiary, portions of the original area are remodelled in this way, having a shaving taken off them as it were and acquiring new and younger superficial deposits.

At no time does the landscape as a whole depart materially from the original planed condition, yet its parts are manifestly of several different "actual" ages,[8] as defined by the various superficial deposits. Unquestionably, the planed landscape is at one point Cretaceous, at another Eocene, at another Pliocene, and at yet another Pleistocene, so that the subjacent bedrock surface is a compound Cretaceous-Eocene-Pliocene-Pleistocene surface. This seems to be essentially the history of the Gobi, and other ancient landscapes *e.g.*, parts of the Kalahari.

But dating land surfaces in this manner is unsatisfactory except for local purposes, and to derive the fundamental or "comparative" age of such a compound landscape we can argue as follows. The landscape, viewed as a whole, was planed first in Mesozoic (Cretaceous)

[8] The age of the surface at any actual spot where datable deposits actually lie upon it.

PLATE 3.—RESIDUAL HILLS IN SOUTHERN AFRICA

FIGURE 1.—Steep-Sided Granite Residuals

Rising abruptly from a smoothly pedimented, erosional plain at the ultimate stage of the cycle of erosion. Eastward from Pietersburg, Transvaal.

FIGURE 2.—Gullied Debris Slope on Loskop, Natal

The change of gradient from debris slope to pediment is clearly shown. T. J. D. Fair photo.

FIGURE 3.—Bari, a Bornhardt 1500 Feet High, Chiweshe Reserve, Southern Rhodesia

The summit of the mountain is closely accordant in level with the Southern Rhodesian plateau, here dissected, though no significant remnant of the plateau surface survives upon the summit. D. Aylen photo.

PLATE 4.—PEDIMENTED LANDSCAPES IN SOUTHERN RHODESIA

FIGURE 1.—Castle Koppies (Tors) Rising Abruptly from Smoothly Concave Pediment

This affords a very clear case of the sharpness of angle which may appear between pediment and free face when there is little waste to form a debris slope. Granite bedrock. Syringa district, Southern Rhodesia.

FIGURE 2.—Broad Pediplain with Steep-Sided Monadnocks

The convexity of the valley sides in the middle distance shows that the stream is newly incised. The landscape is two-cycle. Amapongokwe River, Southern Rhodesia.

FIGURE 3.—Late Stage in Pediplanation Cycle

Residuals rise like islands from the extensive pediplain. Sabi Valley, Southern Rhodesia.

time. From the planed aspect then achieved, it has never since materially departed. What we see now is essentially a Cretaceous surface that has since undergone minor modification only. Its fundamental or "Comparative" age, by which it should be compared with other planed surfaces, is Cretaceous; and so we regard it.

The number and extent of planed Cretaceous land surfaces surviving thus as remnants in landscapes of the present day, especially those of the southern (Gondwana) lands, is remarkable. The survival of these surfaces affords a direct negative to the concept of significant vertical downwearing of landscapes—*i.e.*, to the so-called "Normal Cycle of Erosion".

The rate of retreat of major continental cyclic scarps (*e.g.*, Fig. 3) affords another aspect of the dating of ancient land surfaces. Thus the Natal Drakensberg originated along a coastal flexure which formed the margin of South-East Africa during the Jurassic period (King, 1940; 1944). Wall-like in form, it has retreated over 150 miles westward to its present position. The wall is still 4000 feet or more high with black, forbidding precipices in its upper part along the borders of Basutoland where the crest rises above 10,000 feet. It is quite unmistakable, and shows little sign of flattening in its 120 million years of existence. The rate at which it has migrated averages about a foot in 200 years.

Knowing the date and place of initiation of the scarps and their present position, similar calculations can be made for several other of the world's major cyclic scarps—*e.g.*, the frontier scarp of Moçambique, the scarp of western Mexico, and so forth. Many of these furnish a rate of retreat between 1 foot in 150 years and 1 foot in 300 years. This shows that over long periods, *ceteris paribus*, the rate of scarp retreat is sensibly constant. In other words the erosive processes responsible for such retreat act at much the same rate the world over.

Only when scarps are small (low relief) and the amount of free face and debris slope is much reduced does the rate of retreat slow up. As we have seen, the rate of parallel retreat upon concavo-convex slopes from which these elements are absent appears to be almost negligible.

Quantitative Methods of Landscape Study

However landscapes originate, it is important that their distribution should be known, and for this there is no better method than routine mapping. Erosion-cycle maps, or morphological maps exist of many small areas and even of certain countries—France, Belgium, European Russia, Southern Rhodesia, etc.—but many more such maps are necessary. The data they provide are valuable to engineers, soil conservationists, and others as well as geomorphologists. There is no substitute for such maps, even though useful data may be gleaned from the ordinary topographic sheets.

Methods involving frequency of elevations, spot heights on grids, and so forth have been used (Maze, 1944) to bring out the cyclic facets present in multicyclic landscapes, and several methods involving superimposed and generalized profiles have been employed for similar purposes (*e.g.*, Wooldridge and Morgan, 1937); all have yielded helpful results more or less directly.

The application of statistical analysis is still more recent, and the pioneer has been A. N. Strahler (1950a; 1950b). Statistical analysis is essentially the method of the bulk sample, and is admirable for the study of complex phenomena and processes into which enter a large number of variable factors. As yet few geomorphic topics provide data suited directly to statistical treatment, and methods may have to be adapted to the new field of inquiry, so that too facile results should not be expected. The net result must be, however, a greater precision in our thinking.

At present the method of statistical analysis is in danger of lending an air of truth to erroneous conclusions. Decisions concerning type and admissibility of data need to be made most carefully, or the results will be falsified. No amount of statistical work can improve upon dubious assumptions and we must be ever mindful of Huxley's dictum concerning the use of mathematics (Mathematics is a mill which will grind you meal of any degree of

fineness, but the quality of the meal depends upon the quality of the grain used). Unless the original data and assumptions are above suspicion, the conclusions lack certainty and perhaps validity. Fundamental errors of judgment made in the sampling process are not eliminated in the method, but they may be corrected in the course of further field work.

Strahler (1950a), using valley sides in the Verdugo hills and assuming that they were no longer undercut by stream action, concluded that hillslopes, in general, flatten progressively with the passage of time. Criticisms here would be: (1) The material adduced is too small in bulk to found a general principle upon. (2) It places a relationship between slope angle and stream gradient which may cease to exist after a certain stage is reached in the landscape cycle (*i.e.* slopes may become stabilized in gradient). (3) The manner of slope development is governed by forces active upon the slope and is independent, except initially, of the river. (4) Even upon the slope, erosional agencies do not act uniformly. (5) There are four different slope elements possible, all of which are modified independently. (6) Equilibrium between various features and factors is assumed whereas such equilibrium patently does not necessarily exist in nature. (7) The general conclusion that slopes flatten progressively is denied in nature by the close agreement of maximum slopes in many districts with a mean angle, and the retention of freshness and activity of major cyclic scarps that have existed from mid-Tertiary or even Mesozoic times. This last, under the Euclidean method, reduces the proposition to an absurdity.

Nor can one accept Strahler's general proposition of the equilibrium theory ("decline of stream gradient is accompanied by slope reduction") for it fails to recognize the great diversity of factors operative in the carving of a landscape. No simple relation can exist between hillside slope and thalweg. After the stage of extreme youth they are modified quite independently. What finer examples need we quote than some of the plates reproduced herewith?

To follow the argument further would profit us nothing. Our object here is not the rebuttal of the Equilibrium Theory of Erosional Slopes or any other variant of Davis' early views upon hillslope evolution; it is merely to emphasize that no amount of statistical analysis will compensate for poor geology. Geologists will be grateful for such practical results as will undoubtedly accrue from the use of the new method, but they will not regard it as a substitute for honest field work, nor will they assign to mathematical expressions an importance equal to facts which can be verified and reverified in the field.

Baulig (1950) has made another point: "Engineers will always use more or less empirical formulas for practical purposes. But there remains to be proved that mathematics has ever revealed in geomorphology an actual relationship that had not been discovered without its aid".

What seems necessary now is a multitude of further observations on the nature of landforms, with special reference to hillslopes and perhaps following the lines indicated by Wood, Fair, and Horton (1945). Quantitative work upon pediments and similar topics is needed, and this will lead to statistical studies. An increased volume of measurement upon topographic forms is necessary and should be undertaken directly in the field rather than from maps in the laboratory.

Also required are extensive observations of the processes operating upon landscape, perhaps stemming from the manner of dispersal of rainfall, through water flow to soil creep and weathering. These need to be correlated with different types of climate and terrain. Studies of stream grade, such as those by Kesseli and others, are of less direct application than processes affecting wide areas of landscape simultaneously. Equally required is a multitude of field studies in which cyclic and noncyclic erosion surfaces are mapped in ever-increasing detail. As the number of morphological maps increases, and landscapes are accurately dated, both "actually" and "comparatively", so the erosional histories of the lands will be deciphered and compared, perhaps upon the lines of world-wide correlations.

Canons of Landscape Evolution

Landscape is a function of process, stage, and structure. The relative importance of these is indicated by their order.

(2) The word *epigene* as applied to landscapes means "at the surface" or "subaerial". It does not include landscapes moulded beneath a solid cover of ice, and certain modifications are understood to be necessary in regions of permafrost.

(3) There is a general homology between all epigene landscapes. The differences between landforms of humid-temperate, semiarid, and arid environments are differences only of degree. Thus, for instance, monadnocks and inselbergs are homologous.

(4) Four elements may occur in a hillside slope. From the top, these are: the waxing slope, the free face, the detrital slope, the waning slope (usually pediment). Each or any element may be suppressed on a given hillslope.

(5) Each of the four elements of hillslope may evolve more or less independently, although each affects the others in some degree.

(6) The most active elements of hillslope evolution are the free face and the debris slope. If these are actively eroded, the hillside will retreat parallel to itself.

(7) In planed landscapes, pediments are the most important features. In stable regions like Southern Africa, pediments may occupy more than half the whole landscape, and locally may exceed nine-tenths of the landscape.

(8) The waxing slope is developed under weathering and soil creep.

(9) When the free face and debris slope are inactive, the waxing slope becomes strongly developed and may extend down to meet the waning slope. Such concavo-convex slopes are degenerate.[9]

(10) Parallel retreat of slopes is aided by (a) high relief, tending to maintain a clear free face and a debris slope; (b) resistant formations, tending to make cliffs (a good free face); (c) horizontal structure in sedimentary rocks; (d) generation originally as a tectonic scarp (*e.g.*, fault or monoclinal scarp).

(11) Erosion of the free face and the debris slope is accomplished chiefly by rill wash forming gully heads.

(12) Whereas the transport of debris upon the waxing slope, free face, and detrital slope is governed by both gravity and water work, that upon the pediment is solely accomplished by water work.

(13) Davis' old deduction of continuous lowering of hillside gradients, a feature also of Strahler's "Equilibrium Theory", is incorrect, and never existed as a general process of landscape development apart from terrains of rocks so weak that they cannot maintain a free face and detrital slope.

(14) Rock floors in epigene landscapes appear commonly between the base of hillslopes and stream channels. These rock floors, which are found under all three climatic regimens, originate by retreat of the hillslopes behind them and are subsequently modified by the passage of water across them which confers upon them a concave profile. Such rock floors should, in all cases, be called *pediments*.

(15) Stream spacing is closer in humid than in nonhumid regions so that an evenly distributed rainfall is largely discharged by channel flow. Wider spacing of streams and heavier incidence of rainfall favor sheet flow and also allow room for wider pediments.

(16) Pediments are normally veneered with detrital material which is in process of transport across them. But pediments themselves are essentially cut-rock surfaces.

(17) Pediments which have ceased to evolve may show weathering of the bedrock.

(18) A pediment is the ideal landform for the rapid dispersal of surface water, encouraging sheet flow and with a proper hydraulic profile. Pediments are, indeed, moulded under sheet flow.

(19) The pediment is the fundamental landform to which epigene landscapes tend to be reduced the world over.

(20) Gullying may appear upon pediments where laminar flow of water is changed to linear flow.

(21) The break in profile between pediment and hillside may be abrupt if little detritus is supplied from above.

(22) Quantitative study of both slopes and processes provides a sequence of landscape forms different from those propounded by W. M. Davis.

(23) The early studies in the erosion cycle

[9] Dr. Fair demurs "Rather does the waning slope extend *upwards* to meet the waxing slope and so give the pediment its predominantly concave form". He is thinking chiefly of African conditions, I of European. Both statements may be true.

were conducted in Europe and northeastern North America, both of which regions were previously subjected to a glacial or periglacial climate. These areas and landscapes came to be cited as "normal" for the globe, a misconception that should no longer be tolerated.

(25) The standard or "normal" type of landscape, both now and in the geological past, is the semiarid type with broad pediments and parallel scarp retreat.

(25) Processes of erosion and evolution of landforms can, as a consequence of the above, be best observed in semiarid regions.

(26) Semiarid landscapes are the most efficiently developed. Deviation from the semiarid norm results in less efficient transport of waste, seen on the one hand in the broad, alluviated valley floors and smothered hillslopes of humid regions culminating in moraine and till under glaciation; and the abundant fans and bahadas, or even desert dunes of extremely arid regions.

(27) Water may flow across landscapes either in threads or in sheets. In thin sheets, water may flow in laminae. Such laminar flow is nonturbulent and nonerosive.

(28) In storms of moderate intensity, the manner of water flow appears to best advantage. Rill flow on the steep hillsides is powerfully erosive; in the upper pediment thin laminar flow (nonerosive) may occur, with deeper sheet flood lower down accompanied by turbulence and erosiveness.

(29) Laminar flow past an obstacle shows a depressed water surface, linear flow banks up the water surface against the obstacle.

(30) Only in late Tertiary time have smooth concavo-convex slopes become common. This is a result of retardation of surface wash by a carpet of grass, and consequent enhancement of soil creep.

(31) Before mid-Tertiary time, landscapes generally were of the semiarid (scarp and pediment) type.

(32) Stream work affects the nature of adjacent hillslopes directly only during the early stages of the landscape cycle. After the streams are graded the dominant agencies are the processes acting directly upon the hillslopes, which evolve in an appropriate manner. The streams are, however, affected by the evolution of hillslopes through the nature of the detritus which these supply for transport by the streams.

(33) New cycles may penetrate inland either by nickpoints and incision of the rivers followed by retreat of the valley sides producing "flanking pediments", or by the retreat of wall-like scarps which are independent of the drainage lines "mountain pediments".

(34) On the whole, a landscape is dissected and reduced in a new cycle more rapidly following widespread river incision than by retreat of cyclic scarps originating tectonically. Conversely, the history of a landscape may be deciphered more readily where cyclic surfaces rise steplike between major cyclic scarps than where the landscape has been gutted by stream dissection.

(35) The ultimate cyclic landform is the *pediplain*, consisting dominantly of broad coalescing pediments. Residuals are steep-sided and have concave slopes. Flood plains may or may not be extensive.

(36) A pediplain is multi-concave upward. When the streams begin to incise themselves due to tilting, uplift, or climatic change, convexity enters the landscape adjacent to the stream channels. The interfluves then become transversely convex, and a landform morphologically different from the pediplain is produced. It is an initial stage of the cycle of erosion; the pediplain represents a senile stage. This expresses Penck's *Endrumpf-Primärrumpf* concept insofar as land form is concerned. We do not necessarily follow his further argument relating these differences to decreasing or increasing rates of land movement.

(37) A peneplain in the Davisian sense, resulting from slope reduction and downwearing, does not exist in nature. It should be redefined as "an imaginary landform."

(38) Davis has quoted low granitic domes of arid regions as though their form resulted specifically from long-continued erosion. Observations in Southern Africa show, however, that these forms are normally functions of structure; that, where the vertical systems of jointing are strong, bornhardts and castle koppies (large and small inselbergs) appear, and where flat or gently dipping joint systems are paramount "ruwares" or flat domes of granite appear in the landscape. That such flat domes do not result from the erosion of bornhardts is, for

this region, certain (Pl. 2, fig. 1; Pl. 3, fig. 3); they follow the broadly convex joints formed apparently in plutonic rocks by "unloading" as superincumbent rock systems were removed under erosion.

(39) Monadnocks are concave in profile as a rule (Pl. 4, fig. 3) (including Mt. Monadnock itself) and have originated by surface wash rather than by downweathering and soil creep. They are not necessarily sited upon outcrops of more resistant rocks.

(40) Inherent in the pediplanation cycle, with scarp retreat, are *two* cyclic land surfaces, the older above a retreating scarp, the younger below.

(41) Many of the major cyclic erosion scarps originate tectonically as fault or monoclinal scarps, especially along outwardly tilted coast lines.

(42) Major cyclic erosion scarps retreat almost as fast as the nickpoints which travel up the rivers transversely to the scarp. Such scarps therefore remain essentially linear and do not have very pronounced re-entrants where they cross the rivers.

(43) Major continental erosion scarps in many lands retreat at a rate of about a foot in 150 years to a foot in 300 years.

(44) A landscape once reduced to a pediplain may remain in that state for an indefinite time with only minor alteration, until some change, tectonic or climatic, is introduced.

(45) Notwithstanding the above, small changes continually take place by regrading and conflict of pediments. These changes, insignificant in the landscape as a whole, perhaps amounting to the removal of only a few inches or feet of material, produce great differences in the superficial deposits of the pediments.

(46) Land surfaces may be dated by the deposits upon them. "Actual" ages are local ages fixed by directly dating the deposits in any given locality. "Comparative" ages refer to the dates at which land surfaces were originally bevelled, and are obtained from the oldest superficial deposits.

(47) Land surfaces may bear deposits of any age from the oldest, used for "comparative" dating, to the present day.

(48) A tentative approach, using "comparative" datings, has been made toward the correlation of major cyclic landscapes from continent to continent.

(49) More use of quantitative methods is necessary in landscape study; especially needed is more morphological mapping.

(50) When more suitable data are available, statistical analysis may become a useful tool in landscape study.

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WILCOX DEPOSITS IN EXPLOSION CRATERS, STEWART COUNTY, TENNESSEE, AND THEIR RELATIONS TO ORIGIN AND AGE OF WELLS CREEK BASIN STRUCTURE

BY CHARLES W. WILSON, JR.

ABSTRACT

Four small deposits of Wilcox sediments occur in Stewart County, Tennessee. One of these deposits is in the inner depressed ring, or crater, of the Wells Creek Basin structure. It is concluded that these four craters had a common post-Eutaw, pre-Wilcox age and common origin by the impact and resulting explosions of fragments of a meteor. A 2000-foot core drilled in the center of Wells Creek Basin is summarized.

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INTRODUCTION

Indian Mound Deposit

In about 1930 Dr. Gant Gaither of Hopkinsville, Kentucky, became interested in the commercial possibilities of a local deposit of clay, lignite, and sand located about 3 miles east-northeast of Indian Mound, Stewart County, Tennessee (Fig. 1). Some drilling and opening of pits and shafts was done at that time. As the clay had a high alumina content, the Aluminum Company of America was also interested. During the summer of 1934, the Alcoa Mining Company of New York City prospected the deposit in more detail.

In 1932–1933, Ernest L. Spain, Jr. did field work on this deposit for a Master's thesis for the Department of Geology, Vanderbilt University. The present writer accompanied him during almost all the field work. At that time the pits and shafts were open, and the drill cuttings still available so that a thorough study

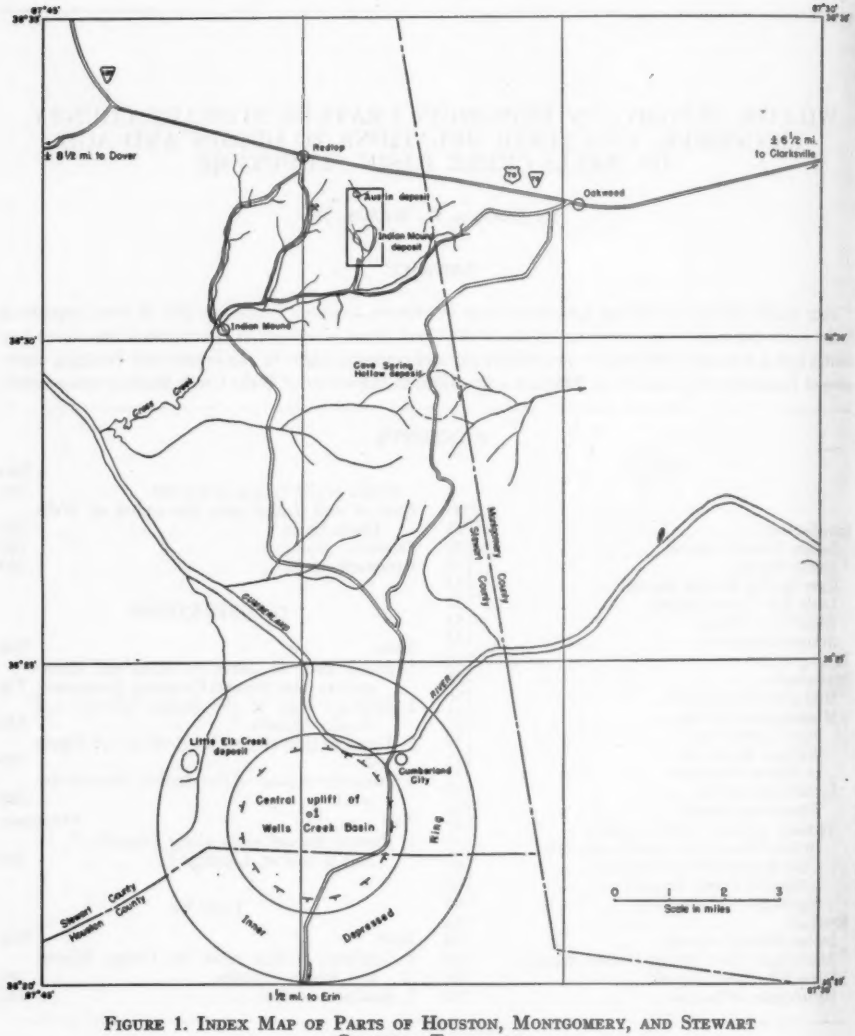

FIGURE 1. INDEX MAP OF PARTS OF HOUSTON, MONTGOMERY, AND STEWART COUNTIES, TENNESSEE

Showing the location and relations of the Wilcox deposits and Wells Creek Basin structure. The rectangular area including the Indian Mound and Austin deposits is the area shown in Figure 2. Locality 1 in the central uplift of Wells Creek Basin is location of the 2000-foot core.

could be made. The thesis was completed by Spain (1933), and copies have been available in the Joint University Libraries and in the Departmental Library of Geology since that time.

Some of the findings of the preliminary exploration were given to Spain by Gaither, but they were insufficient to reveal the full significance of the unique deposit. He concluded that the sand and clay accumulated in ponds and the lignite in swamps along a tributary of Cumberland River, and presented the supposed physiographic changes that permitted such accumulation. He believed that the deposit was of Pleistocene age.

In December 1948, Mr. Lawrence Litchfield,

Jr., of the Alcoa Mining Company made available to the Department of Geology at Vanderbilt its well records, analyses, and location maps of the tract, releasing this information for publication. This information was much more extensive than that available to Spain in 1933 and showed more clearly the character and great thickness of the deposit.

In 1951, Professor Elso S. Barghoorn of Harvard University very kindly studied the microspores of the lignite and concluded that it was post-Cretaceous, pre-Miocene. He also examined the flora of a specimen of lignite from the Wilcox of Graves County, Kentucky. Although the pollen content varied considerably owing to ecological differences, the two have in common the pollen of *Myssa* (gum) and an assortment of pollen of the families Sapotaceae and Fagaceae.

In November 1951 all traces of the exploration had disappeared except for a slight depression representing the shaft (Fig. 2) and a small pile of clay and sand near by. Samples of sand were obtained in April 1952 by hand auger.

As is apparent from the above chronological review of study of the deposit, the writer is deeply indebted to the Alcoa Mining Company, Professor Elso S. Barghoorn, Dr. Gant Gaither, and Ernest L. Spain, Jr.

Austin Deposit

This deposit (Fig. 1), which lies on the Austin property about 1700 feet north of the Indian Mound deposit, was prospected and studied at the same time as the larger one. It is known only in a single well that penetrated 40 feet of transported clay.

Cave Spring Hollow Deposit

A deposit of clay and sand (Fig. 1) is exposed in the head-water area of Cave Spring Hollow 4½ miles south-southeast of Indian Mound and 7 miles north of Cumberland City. It is reported that this deposit was prospected by hand auger at about the same time as the others. The indefinite limits of this deposit (Fig. 1) are based on local reports of where the drilling was concentrated.

Little Elk Creek Deposit

Several small deposits of clay exposed in a tributary of Little Elk Creek (Fig. 1) were reported by Safford in 1869 (p. 349) and were shown to the writer by Dr. W. H. Bucher in about 1933.

Wells Creek Basin

Safford's first geologic map of Tennessee, published in 1856, shows no indication of the Wells Creek Basin structure; presumably, it had not been found at that time. His map of 1869, however, shows the structure, and his *Geology of Tennessee* of the same date discusses the stratigraphy and structure of the area (p. 147, 220, 257, 323, 333).

Dr. W. T. Lander of Williamston, South Carolina, a graduate assistant to Professor Safford at Vanderbilt during the school years of 1887–1888 and 1888–1889, served as Safford's field assistant, during several field seasons, and at some time prepared a detailed manuscript on the annular rings of faults that encircle the central uplift (ca. 1899). At the same time a geologic map and cross sections of the structure were prepared by Safford assisted by Lander. Copies of this map and the cross sections were recently forwarded to Herman W. Ferguson, State Geologist, by the U. S. Geological Survey, and are now available for examination at the Tennessee Division of Geology.

In the period of 1932–1933, Dr. W. H. Bucher mapped the structure for the Tennessee Division of Geology as a project in his study of cryptovolcanic structures. Although his complete manuscript has not been published, a summary of his findings and the geologic map were published in 1933.

The writer studied the stratigraphy of the area for the Division of Geology in 1940. About the same time he made a magnetic map of the region surrounding Wells Creek Basin. This map showed no magnetic anomaly associated with the structure.

Acknowledgments

Acknowledgments have been made to the Alcoa Mining Company, Professor Elso S. Barghoorn, Dr. Gant Gaither, and Ernest L.

Spain, Jr. The writer is also indebted to Dr. W. H. Bucher for a guided tour to the points of interest in the area. Dr. James W. Clarke studied the heavy-mineral content of sand from the Indian Mound, Cave Spring Hollow, and Little Elk Creek deposits. This study strongly supports the correlation of the sand in the deposits.

The Tennessee Division of Geology prepared the analyses of the clay and dolomite included in this report. The Division also purchased aerial photographs of the region between State Highway 76 and Wells Creek Basin in anticipation of locating other deposits of clay. Miss Alice L. Clements of the Division drafted the figures, and Robert E. Hershey studied some of the samples of clay and sand. For these important contributions the writer is indebted to Herman W. Ferguson, State Geologist.

Physiography

The area is located on the Western Highland Rim, which in this part of the State is about 680 feet in altitude. The Rim is locally incised by Cumberland River and its tributaries. The river is at an elevation of about 340 feet. The Indian Mound and Austin area is only a few miles from the river, and the altitude of these deposits is between 460 and 540 feet. The Cave Spring Hollow deposit is about 600 feet above sea level. These deposits of clay do not affect the topography in any way, nor do they show up in the aerial photographs.

As the annular fault rings affect Mississippian limestone and chert of the same resistance, they are not apparent in the topography. The central uplift, however, bringing up older beds of varying resistance, is diagrammatically expressed in the topography. A circular topographic basin has been cut into the older beds to an average elevation of 375 feet. The basin is surrounded by an inward-facing cuesta of resistant Lower Mississippian strata. The annular outcrop belt of the relatively resistant Hermitage formation has been etched into relief of less than 50 feet as a "hogback," and the relatively resistant Knox dolomite and chert form a low rounded hill in the center of the basin, above which it rises about 75 feet. These three topographically high features are separated by an inner "race track," in the less resistant Stones River limestones, and an outer "race track" in the less resistant Silurian and Devonian limestones.

Stratigraphy

Stratigraphic Sequence

The following sequence of formations is exposed in Wells Creek Basin.

	Feet
Mississippian	
St. Louis limestone	200 plus
Warsaw limestone	200 plus
Fort Payne chert	150 plus
Lower Shaly Osage (Ridgetop facies)	50 plus
Devonian	
Chattanooga shale	25 plus
Harriman limestone and chert	50
Ross formation (Birdsong shale and Rockhouse limestone members)	35
Silurian	
Decatur limestone	70
Lobelville formation	30
Bob limestone	25
Beech River formation	40
Dixon formation	40
Lego limestone	30
Waldron shale	Not recognized
Laurel limestone	35
Osgood formation	15
Brassfield limestone	25
Ordovician	
Mannie shale	Present
Fernvale limestone	Present
Hermitage formation	100 plus
Carters limestone	Present
Lebanon limestone	Present
Ridley limestone	Present
Pierce limestone	Not recognized
Murfreesboro limestone	Present
Wells Creek dolomite	Present
"Knox" dolomite	Present

The stratigraphic sequence of this part of the Western Highland Rim and the clay and associated deposits is presented as follows.

Mississippian System

Fort Payne chert.—This chert is a gray or blue-gray, siliceous, calcareous siltstone or siliceous limestone containing small geodes and chert. It weathers to a leached siliceous siltstone and impure, light-brown to yellow chert. Thickness is unknown but is greater than 150 feet. It is not recognized at the surface in the Indian Mound area.

Warsaw limestone.—This massive fine to granular limestone contains some chert and some shaly zones. Many of the blocks of chert that weather from this limestone are fossiliferous and contain small angular holes formed by the leaching of crinoidal fragments, making identification both easy and definite. These blocks are embedded in a matrix of soil or white clay that may be iron-stained locally. This formation forms the lower slopes and the stream valleys in the area of study and is thicker than 200 feet.

In the vicinity of the Indian Mound deposit (Fig. 2), residual clay and chert represent this formation except in the south-central part of the map where Warsaw limestone is well exposed along the creek. A drill hole in the center of the Indian Mound deposit (Fig. 3, A′) penetrated 40 feet of this residual clay and chert. Spain (1933) presents three analyses of the residual clay taken from this well and prepared by D. F. Farrar (Table 2). Note the low average of 0.21 per cent TiO_2 in this residual clay in contrast to the much higher percentages in the transported clay.

St. Louis limestone.—This fine-grained, dense, light-blue to gray, siliceous limestone weathers to a rubble of light-colored chert, often spherical in shape, and silicified fragments of *Lithostrotion.* The rubble caps the interstream ridges at or near the level of the Western Highland Rim. Its thickness is unknown as its top is not preserved in this region, but at least 200 feet is known to be present.

Cretaceous System

Tuscaloosa gravel.—Capping the prominent spurs to the north and west of the Indian Mound and Austin deposits, as well as other ridges in the region, are remnants of the once more extensive blanket of Tuscaloosa gravel. It consists of light-colored rounded chert pebbles and cobbles with some clay. This portion of the Western Highland Rim was once covered with this gravel. It had been largely removed by erosion before the basins in which the clay, lignite, and sand accumulated were formed. While the Indian Mound basin stood open some of the gravel from the surrounding rim rolled into it and was later covered by the Wilcox beds. The gravel was identified in this position in some of the pits at Indian Mound beneath the lignite and other beds.

Tertiary System: Eocene Series

Indian Mound and Austin deposits.—Occupying the large Indian Mound basin (Fig. 2) and the much smaller Austin basin (Fig. 2), which had been formed in the Mississippian and Cretaceous sequence described above, is a complex series of beds and lentils of clay, lignite, and sand (Figs. 3, 4; Pl. 1). The clay and sand replace each other laterally and vertically and are difficult to correlate except in several instances. Locally the two types of sediments grade together. Lignite occurs as a continuous lentil in the area penetrated by holes B-2, B-3, C-1, and C-2. In hole C-2, a lower and thicker lentil of lignite occurs. A small amount of gravel occurs in the sand at depth of 32–39 feet in hole C-2. The clay is locally lignitic.

Spain's (1933) description of the deposit is, as follows:

> "The clay on the Moery farm near Indian Mound occurs as an irregular deposit upon the underlying limestone. It is not a homogeneous deposit, the character of the clay changing both vertically and laterally so that a few feet makes the difference between a smooth, plastic clay containing little grit, and a lens of pure sand. This variation is more the rule toward the surface of the deposit. In the deeper borings the material is more homogeneous except where lignitic layers intervene.
>
> "The sediments in the deposit range from a very fine clay to coarse, gravelly sand. Lenses of micaceous sand, reddish sand coated with iron oxide, pure white sand, lignitic clay, peat, and white clay are interbedded.
>
> "The sands present are rather fine-grained and in some places nearly pure quartz. In other places however, the sands contain an abundance of rutile grains so that they are gray or reddish in appearance. Micaceous sands are present also. The quartz is in very fine grains and the mica is muscovite in the form of tiny flakes. The sand also contains some flaky kaolinite. Some of the sand is stained a reddish or an orange color from a coating of iron oxide on the surface of the grains. This coating can be removed by thorough agitation during washing.
>
> "Although the clays and sands are highly lignitic for as much as 40 feet at times, a layer of pure peat or lignite is rarely found with a thickness of more than a foot or two. True lignite is rarely found, most of the material being peat. Fine sand and clay contained results in a high ash content. An analysis of the ash from a lignite sample showed a TiO_2 content of over 20% that is attributed to the rutile sand present."

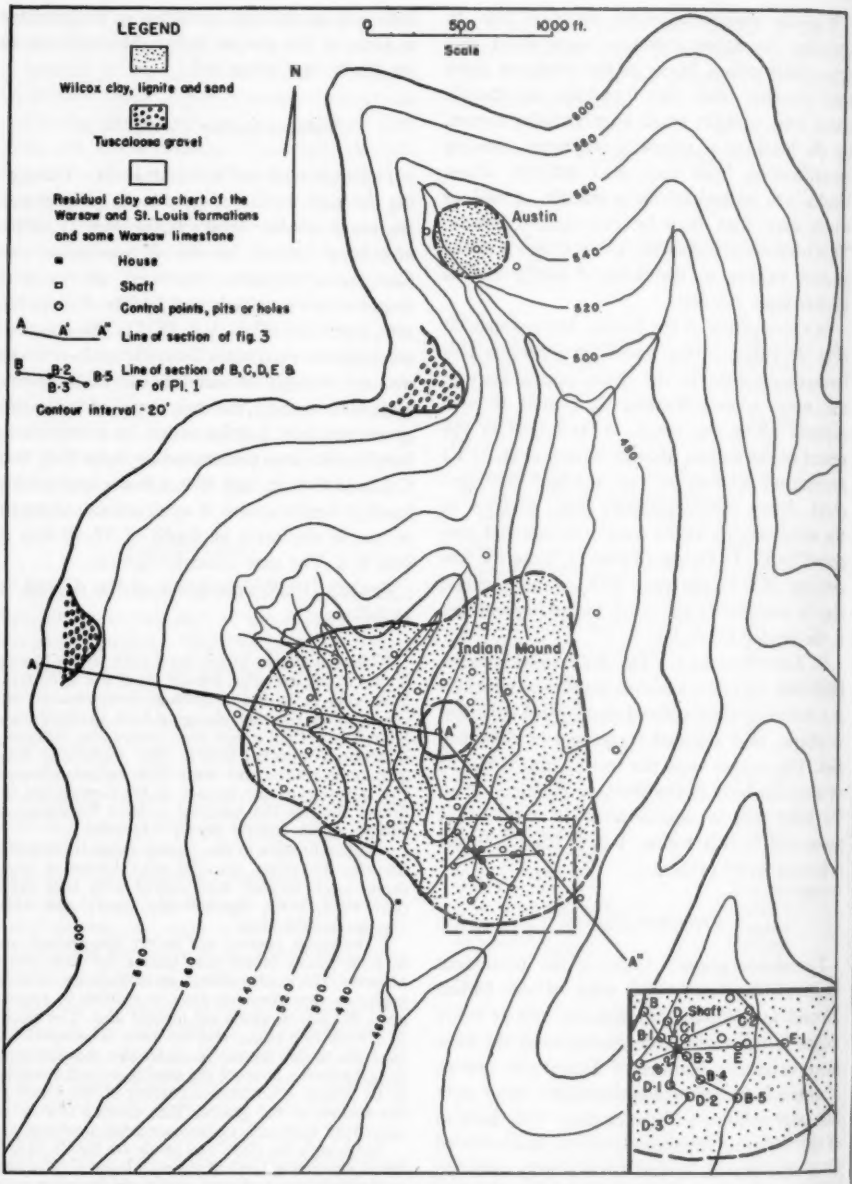

FIGURE 2. GEOLOGIC MAP OF THE INDIAN MOUND AND AUSTIN DEPOSITS

Inset in lower right is an enlargement of the area of the same shape indicated in the southern part of the Indian Mound deposit. This map is modified after Spain (1933).

He described the clay deposits, as follows:

"This variety [transported] of clay is more abundant [than residual clay] and in general overlies the residual clay completely except about the edges of the deposit and on the end of middle ridge. It is made up of fine, detrital material washed into a basin and is interbedded with lignitic layers. In

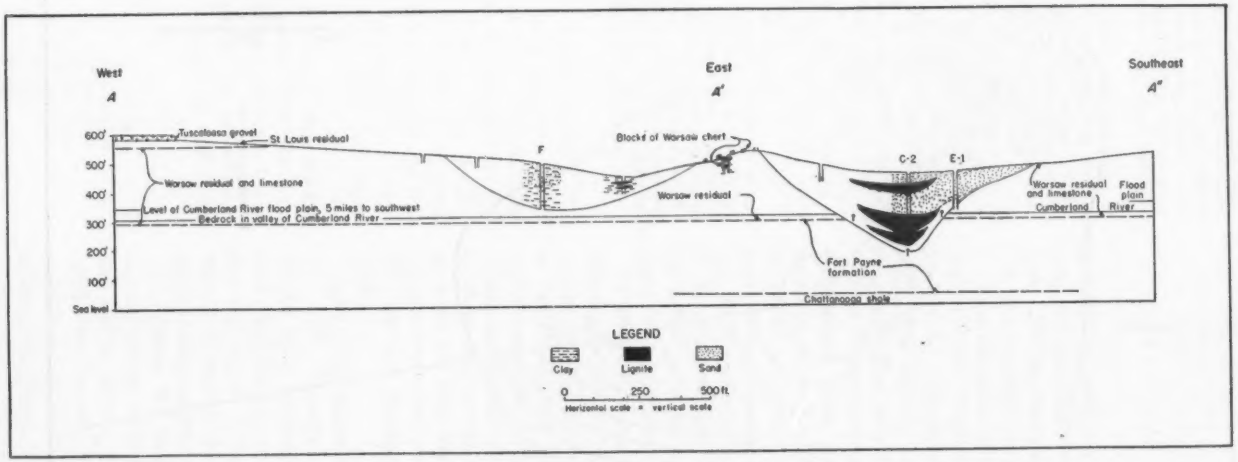

FIGURE 3. CROSS SECTION ALONG LINE A–A′–A″ OF FIGURE 2

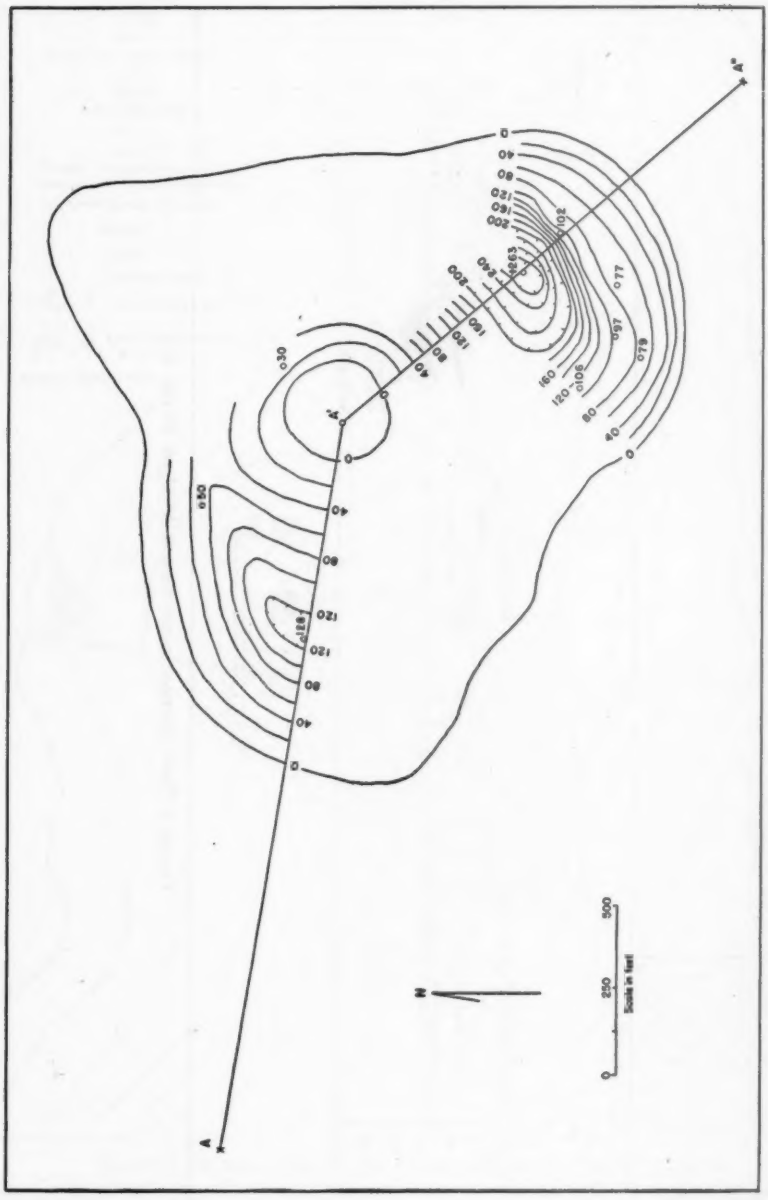

FIGURE 4. ISOPACHOUS MAP OF THE INDIAN MOUND DEPOSIT
Contour interval 20 feet. Well F, as shown on this map, shows 128 feet of deposit, and well C-2 shows more than 263 feet.

places there are thin seams of brown, water-soaked peat which is highly disintegrated so that no recognizable plant remains are to be found. In places the peat has become lignite which is hard and friable and of a dark brown to black color. Interbedded with the clay of white color are clays stained dark by the carbonaceous matter from these seams of organic remains.

"Interbedded with the clays also are lenses of fine grained sand, white or orange in color, some of it micaceous. The sand probably came from the same place as the clay. The test drillings show the clay to be fairly uniform in composition as is shown by the chemical analyses made.

"The clay is of a grayish color when wet but upon drying becomes completely white except the portions containing considerable carbonaceous matter. It is white burning if the content of TiO_2 does not exceed 4 per cent, but if the percentage is too high, the burned clay is yellow in color.

"Fusibility of this clay occurs at cone 13 (1350°C.).

"The presence of titanium in the clay renders it useless for the manufacture of whitewear because it causes uneven shrinkage in burning, resulting in the cracking of the ware. Although the clay burns steel-hard, it is sintered densely by the TiO_2 present [tests by Bureau of Standards].

"The clay has good plasticity, some of it approaching that of a good ball clay. It is smooth and soapy to the touch and contains very little gritty material. The larger fragments left on a 100-mesh sieve consist almost entirely of quartz grains. The clay will pass a 200-mesh sieve leaving a small percentage upon the sieve made up of quartz and rutile with some ilmenite and a very little zircon."

The Indian Mound deposit is about 2000 feet in diameter and has a maximum known depth of more than 263 feet. It is shaped like a doughnut with the central hill of chert occupying the "hole" of the doughnut (Fig. 4). The Austin deposit, about 1700 feet to the north, is no larger than 375 feet in diameter and is at least 40 feet deep, as this thickness of transported clay was penetrated in the hole shown on Figure 2.

The deposit ranges from more than 263 feet thick in well C-2 to a feather edge around the periphery of the deposit and around the border of the encircled central hill. The thicknesses of the Wilcox beds are as follows for each of the key wells:

B	51 feet plus
B-1	55 feet plus
B-2	115 feet plus
B-3	124 feet plus
B-4	55 feet plus
B-5	77 feet
C	106 feet
C-1	64 feet plus
C-2	263 feet plus
D	78 feet plus
D-1	49 feet plus
D-2	97 feet
D-3	79 feet
E	113 feet plus
E-1	102 feet
F	128 feet

In individual wells, the clay has a composite maximum thickness of 98 feet in well B-2; the lignite, 105 feet, and the sand, 123 feet in well C-2. Assuming that the sediments accumulated in a basin formed below the level of the Western Highland Rim, all figures as to thicknesses and depth of the basin represent minima.

The gross lithology of the clay, lignite, and sand is identical with that of the Wilcox beds of West Tennessee about 35 miles west of Indian Mound. Even the dark reddish-brown ironstone concretions that are so common in the Wilcox are present. Analyses (Table 1) and heavy-mineral assemblages also agree in general. More precisely, however, Professor Elso S. Barghoorn studied the microspores in the lignite and concluded that it was post-Cretaceous and pre-Miocene.

Two features of the analyses (Table 1) are noteworthy: (1) At least some of the clay is bauxite, and (2) the analyses have an average of about 6 per cent TiO_2. In the analyses of clay that included TiO_2 from West Tennessee Wilcox deposits the average (Whitlatch, 1940) is 1.5 per cent, and the maximum in any analysis is 2.8 per cent. The occurrence of bauxite has been briefly mentioned by Whitlatch (1939), Thoenen and Burchard (1941), and Bridge (1950).

Spain (1933) reports kaolinite, quartz, muscovite, rutile, ilmenite, and zircon in the clay. Of the heavy minerals, the relative order is rutile, ilmenite, and zircon. In the sand washed out of a sample of clay from Locality F, he reports as much as 30 per cent rutile, the remainder being quartz.

Dr. James W. Clarke studied the heavy-mineral content of three samples of sand from the Indian Mound deposit. Sample No. 1 came from the shaft. Sample No. 2 came from a depth of 9 to 10 feet in a hole 55 feet north of the shaft. Sample No. 3 came from a depth of 10–12 feet in the same hole. His determina-

Contour interval 20 feet. Well F, as shown on this map, shows 128 feet of deposit, and well C-2 shows more than 263 feet.

TABLE 1.—ANALYSES OF CLAY (IN PER CENT) FROM THE INDIAN MOUND AND AUSTIN DEPOSITS

	Depths (Feet)	Al_2O_3	Fe_2O_3	SiO_2	TiO_2	Ignition loss	FeO
PREPARED BY ALCOA MINING COMPANY							
Well B-2							
a	72–74	49.91	0.70	18.16	7.25	23.98	0.10
a	74–78	49.05	0.50	23.24	4.50	22.71	0.10
a	78–82	40.54	0.80	40.33	3.45	14.88	0.10
a	82–86	39.30	1.00	44.16	2.85	12.69	0.10
a	86–92	37.94	1.70	44.97	2.70	12.69	0.10
Well D							
a	42–44	39.02	0.35	44.02	3.30	13.31	0.10
a	44–48	37.58	0.50	40.10	9.35	12.47	0.10
a	48–52	36.32	0.80	38.14	12.75	11.99	0.10
a	52–56	37.43	0.80	40.06	9.70	12.01	0.10
a	56–60	37.08	0.70	40.75	9.55	11.92	0.10
b	71–73	37.68	0.45	35.67	14.00	12.20	0.10
Well D-2							
		Unacct. for					
a	74–76	15.90	1.10	77.60	1.00	4.40	0.43
a	76–80	17.62	1.10	76.93	1.00	3.35	0.22
Well B-3							
		Unacct. for					
a	64–66	50.59	0.85	18.49	6.30	23.77	0.12
a	66–70	54.53	0.65	11.35	5.60	27.87	0.12
a	70–72	53.09	0.65	14.55	5.30	26.21	0.12
b	88–90	49.83	0.90	22.65	6.65	19.97	0.12
b	90–94	57.53	0.75	7.07	6.00	28.65	0.12
b	94–98	50.75	0.80	18.85	5.65	23.95	0.12
c	103–106	40.90	1.05	38.70	2.95	16.40	0.24
Well E							
		Unacct. for					
a	30–32	57.09	0.80	9.67	4.45	27.99	0.10
a	32–36	56.03	1.05	9.91	4.75	28.26	0.10
a	36–39	50.07	1.10	22.05	4.10	22.68	0.10

TABLE 1.—*Cont.*

		Al_2O_3					
PREPARED BY D. F. FARRAR							
Well F							
a	70	31.28	1.35	54.30	1.37		
b	80	29.76		48.90	10.20		
c	115	32.63		45.40	11.30		
d	120	32.69		42.71	7.95		
e	122	29.90		54.86	5.80		
PREPARED BY D. F. FARRAR							
Well in Austin deposit: depth unknown							
		35.58	1.10	46.88	3.60		
PREPARED BY TENNESSEE VALLEY AUTHORITY							
Locality unknown (probably from the shaft)							
		59.40	0.60	4.05	5.80	32.00	
PREPARED BY U. S. GEOLOGICAL SURVEY							
Locality unknown (probably from the shaft)							
		56.80	0.10	6.54	6.38	30.34	

tions of 100 grains from each sample are as follows:

	Sample No. 1	*Sample No. 2*	*Sample No. 3*
Zircon	37	9	4
Ilmenite and leucoxene	33	39	37
Kyanite	13	22	38
Rutile	6	4	2
Epidote	6	10	9
Staurolite	5	12	8
Tourmaline		1	
Pyroxene		1	
Undetermined		2	2

Cave Spring Hollow deposit.—This deposit is located in the headwater area of Cave Spring Hollow 4½ miles south-southeast of Indian Mound and 7 miles north of Cumberland City. Its extent is unknown, and only about 15 feet of clay and sand is exposed. It is reported that the prospecting revealed about 40 feet of clay, but this cannot be checked. If the basin in which this deposit accumulated was formed into the level of the Highland Rim, the deposit would have been thicker and the basin deeper.

The deposit consists of interbedded varicolored clay and sand. One horizon of clay consists largely of flakes of muscovite. The sand contains heavy minerals. In one sample

TABLE 2.—ANALYSES OF CLAY

Residual clay from the center of the Indian Mound Deposit; and transported clay from the Cave Spring Hollow and Little Elk Creek deposits (in per cent)

Depth (Feet)	Fe_2O_3	SiO_2	TiO_2	Ignition loss	Al_2O_3
Well in center of Indian Mound deposit (A′)					
25	?	70.46	0.00	7.16	22.26
25	0.72	90.53	0.55	1.88	5.15
40	0.18	51.16	0.08	12.68	36.26
Clay from the Cave Spring Hollow deposit. Tennessee Division of Geology					
	Al_2O_3	Fe_2O_3	SiO_2	TiO_2	Ignition loss
	9.60	1.75	78.82	1.50	3.46
	9.97	1.28	79.31	0.75	4.49
Clay from the Little Elk Creek deposit. Tennessee Division of Geology					
	6.92	0.24	88.46	0.50	2.44

Clarke made the following count out of 100 grains.

Ilmenite and leucoxene	65
Kyanite	10
Zircon	10
Staurolite and epidote	10
Rutile	5

Analyses of the clay were made by the Tennessee Division of Geology (Table 2).

As this deposit contains no lignite it cannot be dated as successfully as the Indian Mound deposit. The similarity of lithology, heavy minerals, and analyses strongly suggests correlation with the Indian Mound deposit. At one horizon, dark reddish-brown ironstone concretions occur that are identical to those of the Wilcox of West Kentucky. The fact that the two deposits are so close to each other supports their correlation, as deposits of clay and sand of other age are unknown in the region. Pockets of clay do occur in the Tuscaloosa formation near Big Rock in the northern part of Stewart County, but these are very different in appearance and interbedded with gravel.

Little Elk Creek deposit.—In a tributary on the west side of Little Elk Creek 4 miles west of Cumberland City (Fig. 1), several very small deposits of clay are poorly exposed. One of these contains a "clay" composed almost entirely of angular to rounded grains of quartz with flakes of muscovite and heavy minerals. Clarke examined this material and found it to consist of about 99.6 per cent quartz. Out of 100 grains of heavy minerals he found the following concentrations:

Ilmenite (altered in varying amounts to leucoxene)	63
Staurolite	9
Zircon	9
Kyanite	7
Tourmaline	4
Epidote	3
Sillimanite	3
Leucoxene	2

An analysis was prepared by the Division of Geology (Table 2).

The age of this clay is questionable, but it is believed to be the same as the Cave Spring Hollow, Indian Mound, and Austin deposits, for the same reasons cited in the discussion of age of the Cave Spring Hollow deposit. This correlation is strongly supported by the heavy-mineral studies made by Doctor Clarke.

Significance of deposits.—One may assume, correctly or incorrectly, that these four basins, oriented essentially along the same line within a relatively small area, and having received essentially similar sediments with similar suites of heavy minerals, might well have similar age and similar origin. They contain the only such deposits known on the Western Highland Rim.

At the beginning of Wilcox age, four post-Tuscaloosa, or post-Eutaw, basins were present. Transporting streams brought clay and sand from their source area and deposited them in the basins. During intervals of nondeposition of these clastic sediments, swamps occurred, and vegetation accumulated to form the lignite in the Indian Mound basin and possibly in the others. It is believed that Wilcox sediments in the basins were then continuous with those of West Tennessee; if this is correct, the latter extended much farther east than previously postulated.

Reworked Tuscaloosa gravel was found beneath the Wilcox beds at Indian Mound, indicating that it was washed into the basin from

the surrounding remnants of Tuscaloosa before deposition of the Wilcox began. It is surprising that more was not washed in during the deposition of clay and sand. At a depth of 32–39 feet in well C-2, gravel was penetrated, which probably represents reworked Tuscaloosa. The only explanation of the near absence of gravel that the writer can see is that practically all the Tuscaloosa had already been removed from this part of the Highland Rim by the beginning of Wilcox time.

It is very probable that the basins may be dated as post-Eutaw, for the sand of the Eutaw was washed in from the north into the Mississippi Embayment, and the transporting agents would have crossed the same region. If the basins were in existence at that time, they would probably have been filled with sand of this age. The marine Midway of West Tennessee cannot be used in dating the deposits, for it is not known how far east it invaded.

Structure

Indian Mound Deposit

Although correlation of the beds within the Indian Mound deposit (Pl. 1) is not as satisfactory as desired, the writer is firmly convinced from evidence from these wells and from his own examination of the beds in the shaft and pits that these beds were deposited in the Indian Mound basin where they are now preserved and were not let down, as are other isolated deposits of bauxite in the Southeastern States (Bridge, 1950).

No structural deformation in the bedrock of Mississippian age was found, because extreme weathering has reduced the bedrock to a rubble of chert. No evidence of uplift was found, unless the loose blocks of Warsaw chert in the central area of residual chert are higher than their normal position. If the blocks are from the lower part of the Warsaw, then uplift of over 100 feet is possible.

Austin and Cave Spring Hollow Deposits

No structural disturbance was noted in the Austin and Cave Spring Hollow deposits, but again the bedrock is chert rubble yielding no information as to its structure.

Little Elk Creek Deposit

The Little Elk Creek deposit occurs in the inner depressed ring of the Wells Creek Basin structure (Bucher, 1933a). This disturbance is a large explosion structure, with the center only 11 miles south of the Indian Mound deposit (Fig. 1). This structure and the three deposits of clay, etc. to the north lie practically in a straight north-northeast line. Immediately after the deformation there must have been a large crater resulting from the explosion. The clay of this deposit accumulated in this crater.

Significance of Basins

During the post-Eutaw, pre-Wilcox interval, at least four basins were located in this region. The large one to the south, probably 6 miles in diameter, was the crater resulting from the explosion that formed the Wells Creek structure. This crater coincided with the inner depressed ring and contained a central hill, or "spine," coinciding with the area of central uplift. Five miles north-northeast of its northern rim was the much smaller Cave Spring Hollow basin, the extent of which is unknown. Three miles farther north was the Indian Mound basin, at least 2000 feet in diameter and deeper than 263 feet, but with a central hill rising above the level of the floor of the basin. About 1700 feet to the north was the very small Austin basin that was deeper than 40 feet.

It seems logical that the four basins, or craters, had a similar origin at the same time. That origin would have been related to the phenomenon that formed the Wells Creek Basin structure.

Origin of the Basins, or Craters

The origin of Wells Creek Basin structure has been attributed by Bucher (1933a) to a cryptovolcanic explosion, and by Boon and Albritton (1936; 1938a) to a meteoritic explosion. Discussion of the origin of crypto-explosion craters at length is not part of the purpose of this paper, but the meteoritic hypothesis (Boon and Albritton, 1936; 1937; 1938a; 1938b; Baldwin, 1949; Dietz, 1946; 1947) is accepted by the present writer.

The writer believes that the central area of

uplift at the time of explosion was a central hill or "spine" such as found in the Steinheim and Flynn Creek craters, and that the inner depressed ring formed the topographic crater proper with the central hill in the middle. Such a crater would have been 6 miles in diameter. The Little Elk Creek clay and sand were deposited in this crater. All the crater fill except this remnant was later removed by Cumberland River and its tributaries, including Wells Creek. The other deposits to the north are better preserved because of their location in reference to the river. The annular "horsts" and "grabens" would have also been expressed topographically, but not very diagrammatically.

Around the periphery of the central area of uplift the beds dip away from the center as would be expected, but the Ross and Decatur formations dip steeply to the south, or toward the center of uplift, for 1000 feet along the northern periphery as is well seen along the bank of Cumberland River. This is similar to the southward thrust of the Ries Basin in southern Germany (Bucher, 1933b), the southward thrust of Flynn Creek disturbance, Jackson County, Tennessee (Wilson and Born, 1936), and the asymmetry of the Jeptha Knob and Serpent Mound structures (Bucher, 1933a). This asymmetry of Wells Creek Basin, superimposed upon the otherwise circular pattern, has been noted by Bucher (1933a) and Boon and Albritton (1936; 1938a). Lander (Manuscript ca. 1887–1889) and Safford (Map ca. 1887–1889) recognized this bilateral character and, in a sketch in Lander's manuscript, the line of asymmetry is plotted with a strike of N. 25°E. This axis lines up with the northern prolongation of Knox dolomite extending some distance north of the essentially circular outcrop of this formation. The axis and the southward-dipping Ross and Decatur limestone on the northern periphery point unerringly to the Indian Mound crater (Fig. 1).

The writer believes only two known forces could account for the origin of Indian Mound crater: (1) a local, abnormally deep sink hole; (2) the depression ring of an explosion crater. It seems to the writer that the sink hole can be eliminated when, as shown in Figure 3, it must have been cut: (1) 130 feet below the present level of bedrock in Cumberland River valley, and (2) through at least 200 feet of Fort Payne and Ridgetop beds. These relatively insoluble beds are underlain by the Chattanooga shale and about 50 feet of Devonian Harriman chert, a sequence that would have prohibited, or made improbable, the cutting of such a deep sink hole. The writer believes that this basin represents a doughnut-shaped explosion crater with a central hill; it is very similar in cross section to the Steinheim Basin and the Flynn Creek crater. Austin and Cave Spring Hollow craters represent small meteoritic pits, or craters, without known central hills.

Earlier, it was partly proven and partly assumed that the four basins were of the same age and origin. Analyses of the origin of the Indian Mound crater and Wells Creek Basin crater arrive at the same type of origin. Does this not strengthen the assumption of correlation of the clay, etc. deposits?

It is concluded that a swarm of meteors approached the earth's surface from the south, or a single meteor fragmented into at least four pieces before striking the surface. The largest fragment struck at the present position of Wells Creek Basin, and the second in size struck at the Indian Mound locality. Smaller fragments ploughed into the earth to form the Austin and Cave Spring Hollow craters. The Ries Basin and the near-by and much smaller Steinheim Basin are both post-Jurassic, pre-Miocene in age. If these were formed at the same time, which seems logical, they would represent twin craters similar to the Wells Creek Basin and Indian Mound craters.

The meteor struck the surface of the Western Highland Rim capped by St. Louis limestone. Study of the Indian Mound deposit indicates that the Tuscaloosa gravel had been removed from the Rim except for local outliers.

If this picture is correct, then Wells Creek Basin need no longer be dated as post-St. Louis but may now, like the Indian Mound crater, be dated as post-Eutaw, pre-Wilcox, if the Midway sea did not invade this far to the east, or as post-Midway, pre-Wilcox if it did.

Core of Well Drilled Near the Center of Wells Creek Basin

In 1947 the Ordman Company cored Wells Creek Basin on the interpretation that it was

a salt dome, and the core was given to the Tennessee Division of Geology. In December 1951, Robert E. Hershey of the Division and the writer studied the core and prepared the following summary.

The core is essentially complete from a depth of 23 feet to 2000 feet. It started and bottomed in Knox dolomite, which has a great but unknown thickness in this region. There were no solution cavities, and the only portions missing had been jammed in the core barrel and ground up. It is a 3-inch core to about 100 feet, a 2-inch core from there to the bottom. Stylolites are common at all attitudes from vertical to horizontal, are definitely secondary, and occur between rocks of different types or two units of rock of the same type. They are common along the contact of injected breccia and the brecciated blocks of bedrock. Calcite veins are common in limestone and not so common in dolomite.

The injected breccia consists of a matrix of pulverized rock containing fragments of chert, limestone, and dolomite of great variety and usually less than half an inch in maximum dimension. It is much darker than the rock into which it was injected. Although this breccia has been studied, no definite identification of the fragments has been made. The quartz grains and some of the oölitic fragments are probably from the Knox. It is believed that the fragments in the breccia came from many of the formations present in the sequence. Varicolored and waxy-appearing chert that is very abundant throughout all the breccia is identical in appearance to the chert in the St. Louis limestone as exposed in the large quarry about a mile east of Erin. This injected breccia has been found at only one locality at the surface, in the Decatur limestone in the southwestern quadrant of the basin.

One surprising feature of the core is the absence, or near absence, of chert that can definitely be identified as being from the Knox.

As would be expected in a local area of highly porous rock, secondary alteration by ground water is abundantly represented. The calcite veins and vugs, dolomitization, silicification, occurrence of pyrite, manganese dendrites, chalcedonic rings around nuclei, etc., are all of this origin.

Log of core:

(Feet)	
23–143	White to gray completely pulverized rock that contains two thin zones or blocks of fine-grained breccia. Shatter cones at many horizons, particularly well developed at depths of 48 feet and 90–110 feet. The great majority have their long axes horizontal and parallel to each other. Occasionally, the apex may point obliquely up the core. Several vugs contain crystals of calcite, etc.
143–210	Dark-gray fine-grained, very heterogeneous breccia, similar to the injected breccia. Vugs and large shatter cones stop at about 200 feet. Only small, poorly defined and incomplete cones occur deeper except for the one at 1237 feet.
210–274	Light- to dark-gray coarse, less heterogeneous breccia containing fragments of different color and texture cut by dikes of injected breccia.
274–305	Mostly light-gray, very dense dolomite with most of the fragments being from the same beds, demonstrating greater lithological homogeneity. Bedding may be preserved in some of the fragments. Dikes common.
305–315	Mottled gray and brown breccia with its fragments from the same beds.
315–496	Light-gray breccia with larger fragments of fine-grained, dense dolomite, showing still greater homogeneity. Dikes common. At 449 and 488–490 feet dips of about 45° are recognizable. The latter includes beds of sand grains interbedded with sand-free zones. This dip occurs in loose jumbled fragments and has no significance other than that it definitely proves the preservation of bedding planes.
496–657	Alternation of the three preceding types of breccia. Several feet around 615 suggest "flowage" by the elongation of pieces of different color and texture characterized by a wavy appearance.
657–873	Light-gray, fine-grained, very dense limestone, cut by several dikes, one between 683 and 685 feet and another 842–850 feet. This zone shows less deformation and brecciation, although undoubtedly badly sheared. Zone 743–744 is similar to zone 305–315 feet.
873–1067	Essentially the same as the preceding, but more fractured; some of the breccia is heterogeneous. Several inches of green shale at 911 feet have

dips of 50°. Between 1025 and 1057 feet, the core is white, unfractured bedrock. At 1063 and 1067 there are 1-foot zones of calcareous sandy beds.

1067–1743 White to gray unbrecciated bedrock, only locally brecciated. Below 1067, more than 10 feet of the core shows dip of 45° that may represent to some extent the first dip of bedrock. An oblique shatter cone with 45-degree angle to the vertical core was found at 1237 feet. Oölitic zones occur between 1317 and 1327 and 1353 and 1363. The frequency and thickness of dikes are greatly reduced below 1067 feet with several intervals of as much as 100 feet having no dikes. The great majority of those present are ¼–½ inch thick. Between 1554 and 1557 feet a vertical dike, up to ½ inch thick tapers downward, but leaves the core at 1557 before pinching out. Deformation is definitely greatly reduced in this interval.

1743–1930 This part of the core shows deeper brecciation and injection of dikes. The breccia is both heterogeneous and homogeneous. There are many dikes between 1786 and 1799. The heterogeneous breccia disappears downward; in the lower part only homogeneous breccias are present. Dikes also decrease.

1930–1989 Homogeneous breccia, with no dikes between 1942 and 1989 feet, where there is a 2-inch dike, the lowest in the core. This dike is much lighter in color and contains larger fragments of the bedrock than any of the rest.

1989–2000 Sheared, but unbrecciated bedrock.

Analyses of the upper 227 feet of the dolomite were prepared by the Division of Geology. The analyses varied but little in the 17 samples analyzed. The estimated average analysis, prepared by D. F. Farrar, is as follows:

Insoluble and silica, 6%; Iron oxide and alumina, 1%; $CaCO_3$, 58%; $MgCO_3$, 35%.

The examination of this core was an unusual privilege and in a way an eerie experience. The deep fingers of grotesque injection dikes and the intense, bizarre, ever-changing patterns of brecciation and deformation are awe-inspiring. Each new box of cores revealed new, strange, and different intricacies.

This core presents three significant features. (1) Deformation was instantaneous, and did not result from normal tectonic stresses; (2) progressive downward dying out of deformation[1] may be traced, in spite of the brecciation between 1743 and 1930 feet; (3) in the top 200 feet of the core, the shatter cones are all horizontal, except for some that point obliquely upward. The greatest concentration of horizontal cones is at a depth of 100 feet. Below 200 feet, shatter cones are rare, incomplete, and poorly defined with a single exception at 1237 feet. As the core was not oriented, it is impossible to state in which direction these cones pointed. In an exposure about 650 feet south of the well, shatter cones are horizontal, and their axes strike N. 7° W. They occur in a vertical block of dolomite with strike of N. 20° E. These cones were not formed by the impact of the meteorite, as such should be normal to the bedding and oriented stratigraphically up, but rather by the explosion of the rocks compressed beneath the penetrating meteorite. This block must have been knocked out of position by impact and penetration a few moments before the explosion. It is believed that the cones in the core pointed northward as do those in the near-by exposure. At a depth of

[1] On April 25, 1946, the writer received the following statement from Mr. W. B. Weeks of the Phillips Petroleum Company in regard to two wells drilled on the Sierra Madera structure, Pecos County, Texas. This structure had the same origin as the Wells Creek Basin structure, and dips at the surface are very steep, ranging up to vertical.

"The hole was abandoned March 15, 1946, at a total depth of 12,096 feet in lower Pennsylvanian strata of possible Strawn age composed of black shale with beds of limestone. Leonard beds from the surface to 4400 feet were composed chiefly of dolomitic limestone. Wolfcamp beds, 4400 to approximately 5535 feet were composed of interbedded limestone, dolomitic limestone, and shale. Strata below that depth, probably Pennsylvanian, were composed largely of shale with beds of sandstone and occasional limestone. There was an increase in the amount of limestone below 10,300 feet.

"No volcanic type rocks were encountered and there was no evidence of mineralization or metamorphism to indicate nearby igneous intrusives. A core 8110-16 recovered 6 feet of interbedded argillaceous sand and dark shale with bedding planes indicating dips of from 5 to 15 degrees. No other information on dips was obtained.

"The old Thompson No. 1 Elsinore well, less than one-half mile to the northeast and down the slope of the mountain, was drilled only to a depth of 7212 feet. A tentative correlation places it approximately 470 feet lower structurally than the deeper hole."

1237 feet, an oblique shatter cone points upward at a 45 degree dip. Its orientation is likewise unknown, but, from the evidence furnished by the exposure, it is believed to have been the same. The orientation of the shatter cones in the exposure and the postulated similar orientation of those in the core suggest that the center of explosion was north of the well.

These features are believed to present definite evidence that the deformative force came from above and not from below. This evidence combined with the occurrence of four aligned craters, of which the Indian Mound crater has critical depth and cross section, and the southward dip of the Ross and Decatur limestones on the north periphery of the uplift of Wells Creek Basin all harmonize to tell the same story of meteoritic origin.

Economic Geology

At least a brief statement of the economic aspects should be presented. The analyses of bauxite, clay, dolomite, etc. (Tables 1 and 2) supplement this statement. In the Indian Mound crater, the exceedingly thick lignite encountered in well C-2 is a possibility. The plastic clay, clay with high rutile content, clay with high muscovite content, bauxite, and rutile are also available. The dolomite at the surface of Wells Creek Basin may be considered an important mineral resource. The brecciated dolomite in the core demonstrates that a beautiful dimension stone "marble" is present.

All these resources are either small in quantity or too deeply buried, with the exception of the dolomite, but nevertheless they are there. Future needs could make any one of them commercial.

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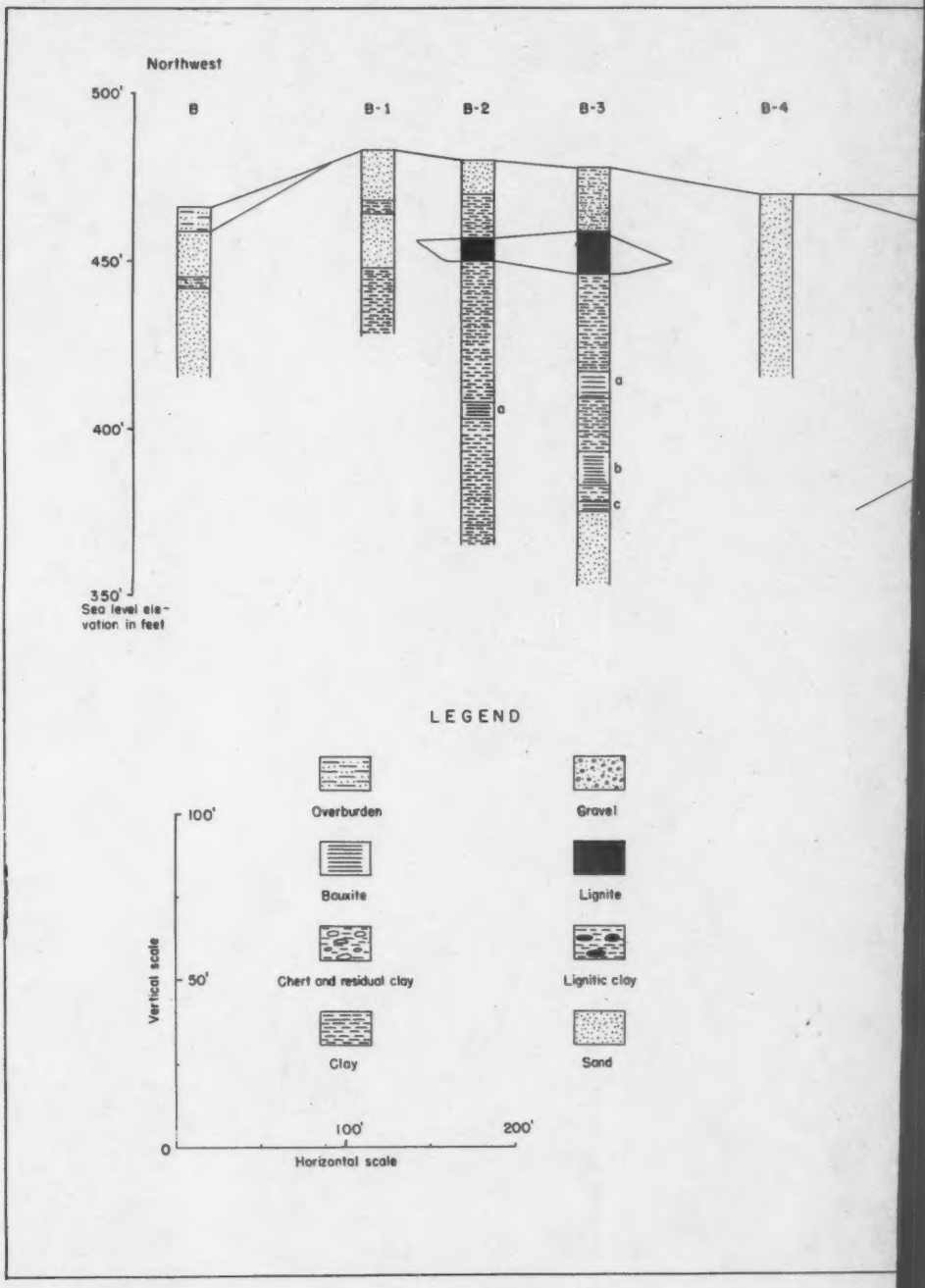
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Lignite
Chert and residual clay
Lignitic clay
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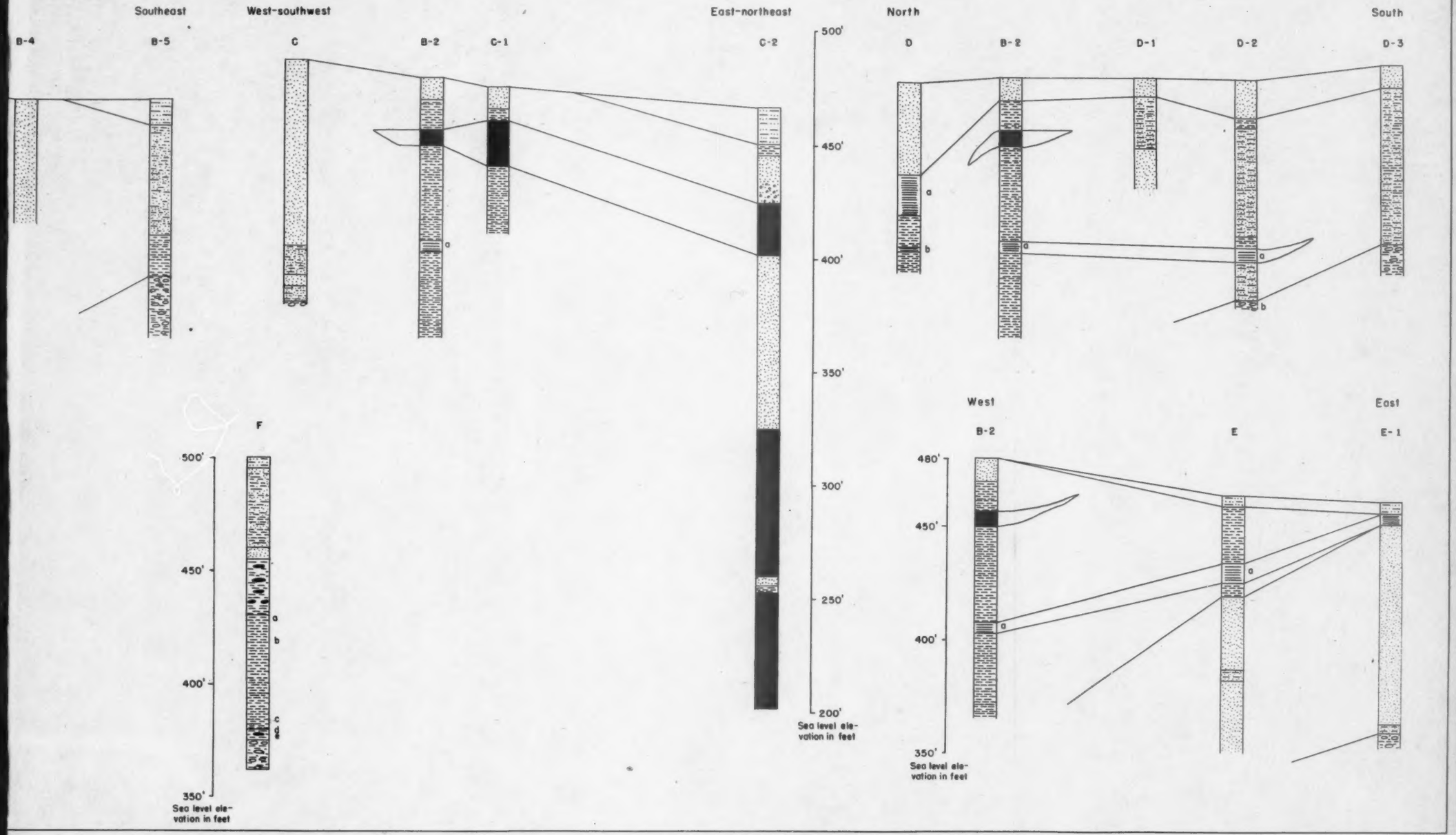
200'
Horizontal scale



PLOTTED LOGS OF WELLS ALONG LINES B, C, D, AND E AND AT LOCALITY F

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PETROCHEMICAL STUDIES ON THE ALKALIC ROCKS OF THE MOROTU DISTRICT, SAKHALIN

By Kenzo Yagi

Abstract

The alkalic rocks of the Morotu district occur as sheets, laccoliths, and dikes in the Tertiary formations and are probably Pliocene. The rocks vary from dolerite, the most abundant, through monzonite to syenite. Most of the intrusive bodies, except small sheets and dikes, are composed of dolerite in the marginal parts and monzonite or syenite or both in the central parts, with gradational boundaries. Monzonite and syenite also occur as irregular veinlets or schlieren in the doleritic parts, with sharp boundaries. Thus differentiation *in situ* is strikingly displayed. These intrusive bodies are, therefore, formed by composite intrusion. Contact effects of the intrusions on the Tertiary formations are weak and are confined to the immediate contact zone. Phenomena of assimilation have not been observed.

Principal minerals of these rocks are plagioclase, anorthoclase, microperthite, olivine, titan- and alkali-pyroxenes, alkali-amphiboles, titanbiotite, analcite, and iron ores. Feldspar evolves from labradorite to oligoclase and is joined by anorthoclase and microperthite in the later stage. The trend is always toward enrichment in the Or molecule. Pyroxene, the most important mafic mineral, forms a continuous reaction series from diopsidic augite through titanaugite, soda-augite, and aegirinaugite to aegirine. Amphibole also forms a reaction series from barkevikite through kaersutite and hastingsite to arfvedsonite, but it is not clear whether their relation is continuous or discontinuous. Biotite is abundant in the dolerites, but very low in the syenites. Analcite is abundant in all rocks, apparently formed hydrothermally in the later stages.

Chemically these rocks range from 46 to 61 per cent SiO_2. Their alkali-lime index is 50.0, placing the suite in the "alkalic series." The parental magma is inferred to be olivine basaltic magma with moderately alkalic affinity. The various rock types are believed to have been formed chiefly by crystallization differentiation from the parental magma after it intruded into the present position. Volatile components have also played an important role during crystallization, especially in the later stage.

There is no absolute enrichment in iron throughout the crystallization course of the Morotu rocks. It is inferred that not only the shape of the intrusive bodies but also the composition of the parental magma determine whether iron enrichment or alkali and silica enrichment will prevail in the intrusive bodies.

Its geographical position, its period of activity, and its petrological character justify the conclusion that the Morotu district constitutes the northeastern end of the Circum-Japan Sea province of Cenozoic alkalic rocks.

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Introduction

Since about 1910, petrologists have noted the occurrence of alkalic rocks in Japan and vicinity and have generally accepted the existence of a petrographic province of alkalic rocks around the Japan Sea. Korea, especially, provides many localities of feldspathoidal rocks, and the Oki Islands of Japan exhibit a remarkable suite of nonfeldspathoidal alkalic rocks. However, the northern part of this Circum-Japan Sea alkalic province has not received much attention. In the course of his researches on the coal field of Southern Sakhalin, J. Takahashi found peculiar occurrences of alkalic rocks in the Morotu and Nayosi districts on the western coast, and alkalic rocks have since been described by Ishibashi (1937, p. 244) and by Iwao (1939, p. 155) with special reference to the mineral paragenesis.

At Doctor Takahashi's suggestion, the writer made a field survey in the Morotu district in the summer of 1944 for about a month. Hitherto no chemical data were available for these alkalic rocks, and, in view of the importance of the chemical composition of the rock-forming minerals for the petrogenetical studies, the writer has made chemical analyses of nearly 20 minerals separated from the parent rocks, as well as analyses of many rock types. With these chemical data the writer presents a quantitative picture of the evolution of the alkalic rocks in the district.

Many field problems and some of the interpretations remain incomplete because it is difficult to revisit the district since its restoration to the U.S.S.R. after World War II. However, in the hope of contributing to our knowledge of the alkalic rocks, the writer submits the results of his research.

Acknowledgments

The writer takes pleasure in expressing his sincere thanks to Dr. Junichi Takahashi, President of Sinsyu University and former Professor at the Tohoku University, for his kind guidance, valuable advice, and criticism throughout the study. He is also greatly indebted to Professor Manjiro Watanabe of the Tohoku University

FIGURE 1. INDEX MAP OF THE MOROTU DISTRICT, SAKHALIN, AND OTHER LOCALITIES OF THE CIRCUM-JAPAN SEA ALKALIC PROVINCE

M	Morotu	Me	Meisen
A	Atumi	N	Nagahama
Ab	Abekawa	Ne	Nemuro
H	Hakuto	O	Oki
K	Kyusyu	S	Saisyu Island
Ka	Kainei	U	Uturyo Island

for suggestions. The main part of the work was done in the Petrological Institute of the Tohoku University, and the rest in the Geophysical Laboratory through the courtesy of Dr. L. H. Adams, Director of the Laboratory. The writer is grateful to Drs. L. H. Adams, N. L. Bowen, and F. Chayes of the Geophysical Laboratory and to Dr. S. S. Goldich of the University of Minnesota for advice and criticism. Thanks are also due Messrs. T. Koiwai, Koshitani, Ogasawara, and Kannari for their courtesies during the field survey, and Dr. S. Iwao and Mr. J. Masui for their kindness in placing some of the rock specimens at the writer's disposal.

GEOLOGICAL OBSERVATIONS

Location and Physiography

The Morotu district is on the western coast of Sakhalin, about 70 km. south of the 50° N. parallel, which was the boundary between the Japanese and the Russian territories before 1945[1] (Fig. 1). The area covered by this study is about 15 km. north-south and 10 km. east-west. On the western coast is the Morotu coal mine, for which the district is named. Just north

[1] The Japanese place names are used in this paper. They probably are now changed to Russian names, but the writer has had no opportunity to ascertain these.

of the district is Nayosi, a harbor and the commercial center of this area.

The district is mostly low mountainous land. The highest peaks, Mt. Morotu and Mt. South Morotu, each more than 700 m. in elevation, are in the center of that area. The mountains lie in several rows trending northwest-southeast, roughly parallel to each other. The chief rivers, the Kitakozawa, the Morotu, and the Tiyo, all flow from southeast to northwest and empty into the Mamiya Strait.

TABLE 1.—STRATIGRAPHIC SUCCESSION OF THE WESTERN COASTAL REGION OF SAKHALIN

(TAKAHASHI, 1942)

Formation	Rocks	Age
Onnai	Shale, sandstone	Pliocene-Miocene
Nayosi coal-bearing	Alternation of shale and sandstone	Miocene
Akusyu	Volcanic rocks and their pyroclastics	Miocene
Nisisakotan	Shale	Miocene
Yokunai	Alternation of shale and sandstone	Miocene-Oligocene
—	Unconformity	—
"Upper Cretaceous"	Green sandstone, shale and marl	Cretaceous
"Lower Cretaceous"	Shale and sandstone	Cretaceous

The development of the coastal terraces is remarkable. Those 20 and 40 m. above sea level are most prominent and have well-preserved initial surfaces. Those 80 m. and 120 m. high are more or less dissected and are sometimes separated into dome-shaped hills. As the area is generally covered by thick forests of fir and pine, exposures are usually poor, and the relations of rocks can be observed only in the cliffs along the coast or the rivers.

Geological Setting

The western coast of Sakhalin is composed mainly of Tertiary and Cretaceous formations. The stratigraphy of this region is shown in Table 1 (Takahashi, 1942, p. 59). In the Morotu district, only the Tertiary, Onnai, Nayosi, and Akusyu formations are developed (Fig. 2). The Nayosi formation, which covers more than half the district, contains numerous excellent coal seams that are being worked. The Onnai formation, resting conformably on the Nayosi, is confined to the synclinal part. The two formations were intruded by the alkalic igneous rocks with only slight metamorphic effects on the sediments. The Akusyu formation forms the core of the anticlinal zone, and exposures are very limited. It is chiefly composed of lavas of andesite and dacite and related tuff or tuff-breccia. The alkalic rocks rarely intrude these rocks. The andesite is composed of phenocrysts of labradorite (An_{55}) and augite in a groundmass of andesine (An_{40}), granular augite, and magnetite. Aggregates of chlorite and calcite are common. The dacite has many phenocrysts of quartz and albitized plagioclase in a groundmass composed of quartz, plagioclase, chlorite, and devitrified glass. Most of the colored minerals are altered, and their original composition is unknown. Although all the volcanic rocks of the Akusyu formation are more or less altered, it is evident from their mineralogical composition that they are all calcalkalic rocks.

As the alkalic igneous rocks are intrusive into the Onnai shale, their age is post-Onnai, or probably Pliocene. It is worth noting that the igneous activity of this region once produced calcalkalic suites and, later, alkalic rocks. However, no evidence of the genetic relation between the two series has been observed in the field.

Modes of Occurrence of the Alkalic Rocks

The alkalic rocks form dikes, sheets, or laccoliths of various sizes in the Tertiary formations. The marginal parts of these intrusive bodies frequently are composed of dark dolerites, and the central parts of leucocratic monzonite or syenite. Field observations on some occurrences are given below.

MOROTU CAPE: The northern part of the cape is composed of the Onnai shale, striking N. 40°–50° W. and dipping up to 20° NE., and of the dolerite intruded concordantly into the shale (Fig. 3). The dolerite shows remark-

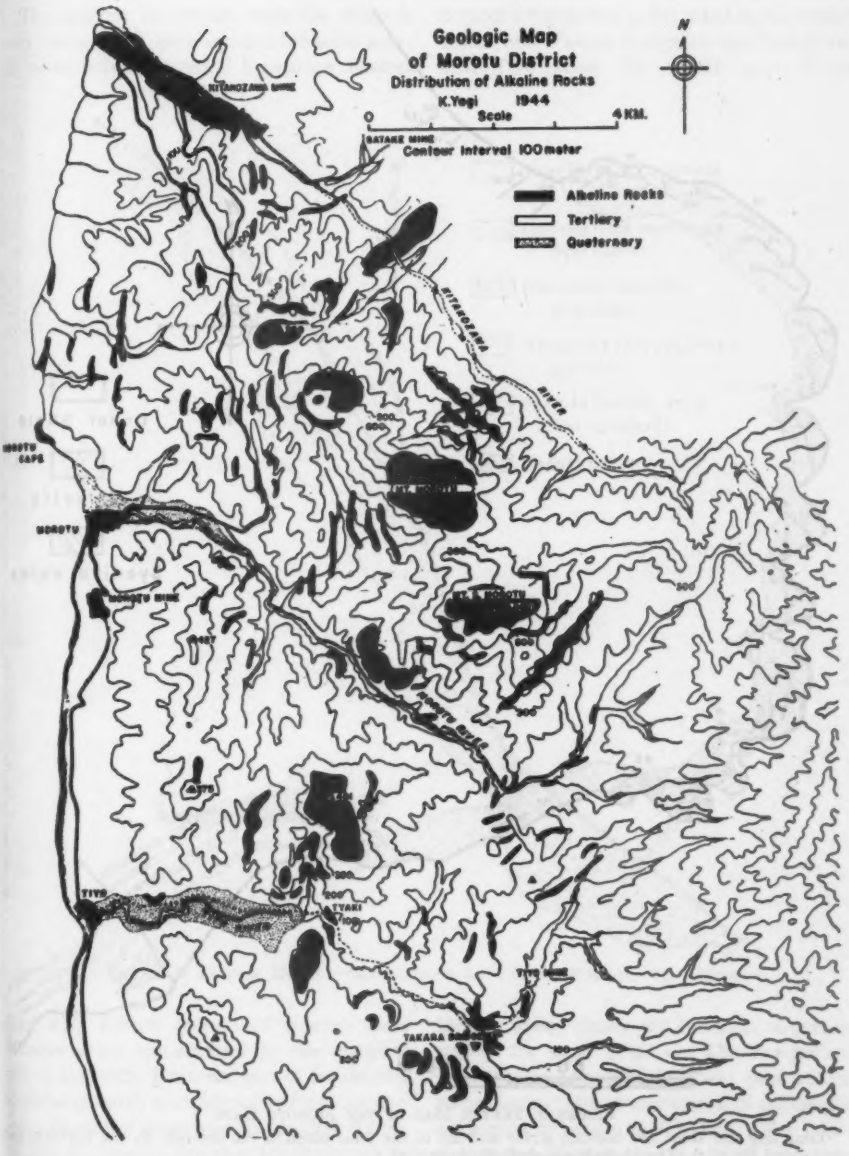

FIGURE 2. GEOLOGIC MAP OF THE MOROTU DISTRICT

able jointing, with a nearly flat-lying system most prominent. The strike of the joint planes swings and is roughly parallel to the shore with gentle dips outward (10°–20° W.), but coincides with those of the shale in the northern part. The contact between dolerite and shale is parallel to the stratification in the northern part of the cape, but is transgressive in the southern part. Therefore the intrusive body is considered a laccolith in the shale, disregarding

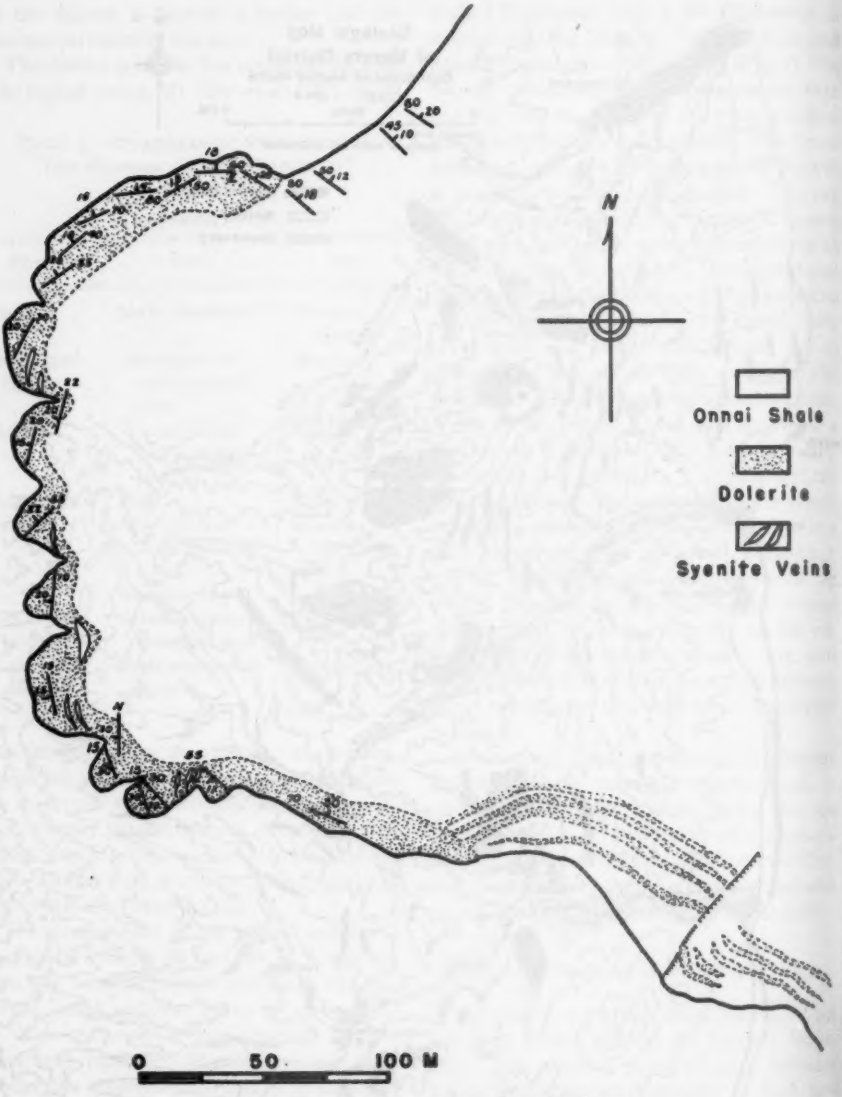

FIGURE 3. SKETCH MAP OF THE MOROTU CAPE

Long line and short bar indicate strike and dip of the joint plane in the dolerite. In the northern part strike and dip of the Onnai shale are similarly indicated.

minor discordant relations. In the southeastern part of the cape there is a swarm of parallel dikes.

The dolerite is compact and fine-grained near the margin, and coarse-grained in the central part. Shale more than 50 cm. from the contact surface is free from the metamorphic effects, but nearer the contact it is compact—sometimes even flinty—and bleached, owing to the volatilization of organic matter. Many minute calcite crystals are formed in such places.

The dolerite in contact with the shale is very fine-grained, compact, and brownish gray. It is sometimes traversed by narrow fissures lined with minute crystals of quartz. These siliceous veins are confined to the marginal part of the body. The inner part of the dolerite is coarse-grained, sometimes dioritic in appearance. There are many syenitic veins 1–50 cm. thick in the dolerite, and they are either regular networks, irregular veinlets, or schlieren. The schlieren are very coarse-grained and contain long prismatic crystals of augite up to 5 cm. long, which are arranged normal to the boundary surface.

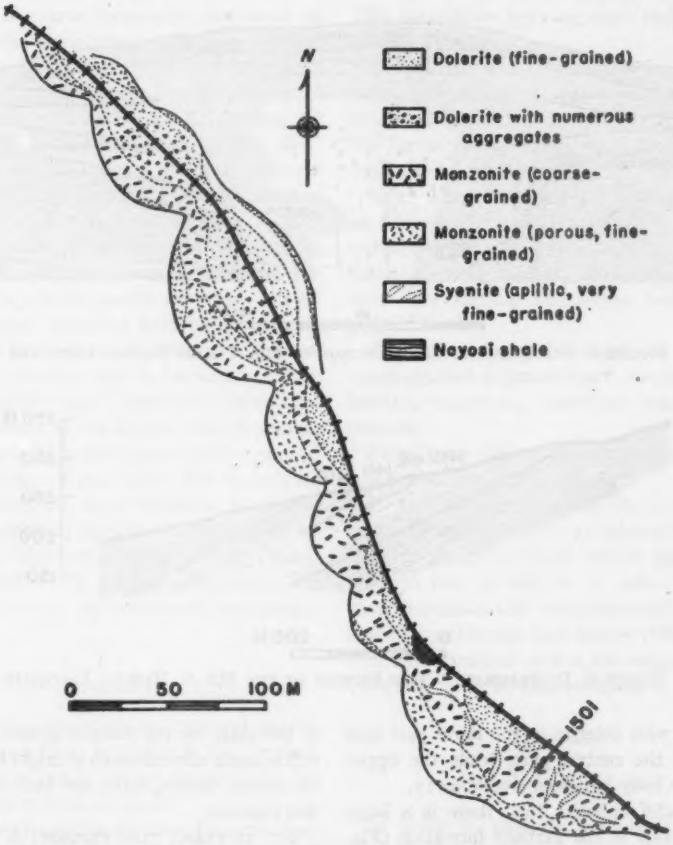

FIGURE 4. SKETCH MAP OF THE VICINITY OF TAKARA BRIDGE, TIYO MINE

TAKARA BRIDGE, TIYO COAL MINE: Good exposures are observed in the cut along the trolley tracks near Takara Bridge on the Tiyo River, near Tiyo Mine. The sketch map of this locality shows clearly the variation of texture within the mass (Fig. 4). The fine-grained dolerites are found in the lower part and in some places contain numerous small pea-shaped aggregates of augite and feldspar. The upper part of the cliff shows coarse-grained monzonite which contains some stellate aggregates of long prismatic crystals of augite. In the uppermost part is found rather fine-grained monzonite. These rock types grade into each other without sharp boundaries. Numerous veinlets of syenite traverse the other rock types, always

with sharp contact. Slightly metamorphosed shale is found in the lower part in contact with the fine-grained dolerite. These facts suggest that the intrusive body also forms a sheet in within the intrusive body. The shale near the base is in some places brecciated and mingled with fragments of dolerite. This is probably the neck of the laccolith. Evidence of assimilation

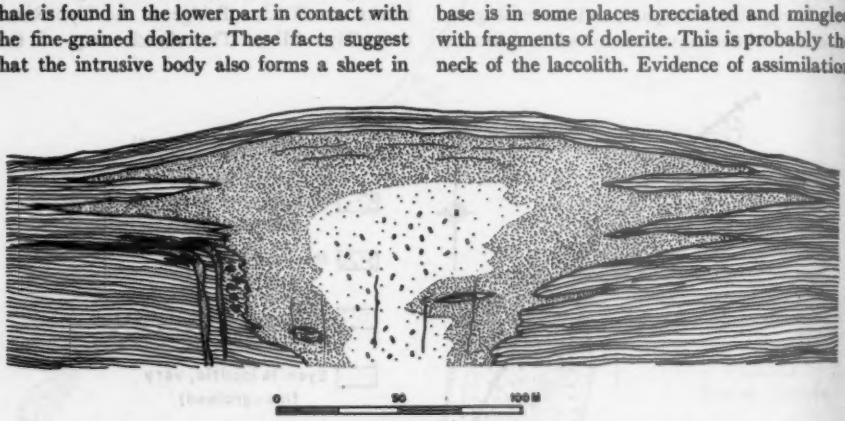

FIGURE 5. DIAGRAMMATIC CROSS SECTION OF THE TAKARA BRIDGE LACCOLITH

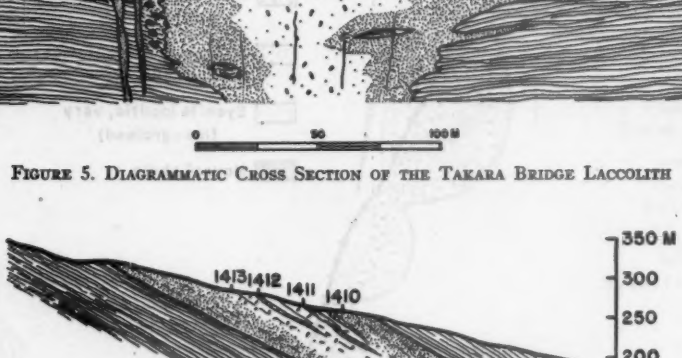

FIGURE 6. DIAGRAMMATIC CROSS SECTION OF THE MT. S. MOROTU LACCOLITH

the shale, with dolerite in the lower part and monzonite the central part, while the upper half of the body has been eroded away.

In the cliff opposite this, there is a large intrusive body in the Tertiary formation (Fig. 5). The body is a typical laccolith of cedar-tree type with flat top and nearly horizontal lateral offshoots. It is about 90 m. high and more than 200 m. wide. Bleached and hardened shale lies immediately on the fine-grained dolerite. The dolerite has platy joints nearly parallel to the upper boundary and to the stratification of the shale. The offshoots of sheets also are fine-grained dolerite. The rock is coarser and more leucocratic toward the inner parts and in the central part is composed of monzonite traversed by many irregular schlieren or veins of syenite. Coarse-grained dolerite appears again in the lower part and is especially well developed around the large xenolithic fragments of shale of the shale by the dolerite is lacking. Layers rich in mafic minerals such as might be produced by crystal settling have not been observed in this exposure.

CUT AT TYAKI, TIYO VILLAGE: A thick sheet found in the cut at Tyaki, about 4 km. east of Tiyo Village, is composed mostly of weathered dolerite. It shows onion structure following the cuboidal joints. Numerous pea-shaped aggregates of augite and feldspar are found in the fresh dolerite. They are more conspicuous in the decomposed dolerite, owing to their greater resistance to weathering. Irregular veins or schlieren of monzonite, 50 cm. to 5 m. thick, are sporadically distributed in the dolerite. The monzonite is coarse-grained and remains fresh in contrast to the easily decomposed surrounding dolerite.

LACCOLITH OF MT. SOUTH MOROTU: Various kinds of alkalic rocks as well as metamor-

phosed shale and sandstone are exposed in a small ravine about 1 km. southeast of Mt. South Morotu. The upper and lower parts are fine- or medium-grained dolerite, and the central part is coarse monzonite, traversed by many coarse, sometimes pegmatitic, veins of syenite. The dolerite and monzonite have platy joints nearly parallel to the stratification of the Tertiary beds striking N. 40° E., dipping 20° SE. The rocks form a thick laccolith (Fig. 6).

KITAKOZAWA COAL MINE AND ITS VICINITY: Contact action by the dolerite on coal beds was observed at the Kitakozawa coal mine. Natural coke and anthracite were formed according to the various grades of metamorphism by the dolerite. The dolerite itself is also highly altered in such cases, becoming pale bluish gray or brown, and is traversed by numerous veins of calcite. In extreme cases it becomes a clayey material, called "don," mixed with carbonates.

A small sheet at the Satake Mine southeast of Kitakozawa Mine has a border of fine-grained dolerite, about 10 cm. thick. The underlying zone, of about the same thickness, is dolerite with amygdaloidal cavities filled with calcite or analcite. This zone grades to very coarse-grained dolerite with diabasic appearance. The sheet has, however, neither syenitic nor monzonitic facies.

Summary

The alkalic igneous rocks occur as sheets, laccoliths, and dikes intrusive in the Tertiary formations, especially in the Onnai shale. No lava flows have been observed.

Sheets and laccoliths are more frequent than dikes and favor the central portion of anticlinal folding or domes in the Tertiary formations.

Contact effects by the intrusive bodies on the invaded Tertiary formations are confined to the vicinity of the immediate contact and are always weak.

Neither assimilation nor hybrid phenomena are observed between the intrusive bodies and the country rocks, although xenolithic fragments of the sediments are rare in the dolerite.

Small dikes or thin sheets less than 1 m. thick show no appreciable variation in texture and mineral composition from margin to core. They usually consist of uniform dolerite.

The thicker sheets and laccoliths are composed of different rock types. The marginal parts are dolerite, either fine- or coarse-grained. Joints parallel to the contact boundaries are sometimes well developed in such parts. The inner parts are monzonite, syenite, or both. The boundaries between each rock type are usually gradual.

Monzonites and syenites also form irregular veins and schlieren with sharp boundaries. They are never found intruding directly into the Tertiary formations but are always confined within the doleritic parts.

Even where the boundary is sharp, there is no difference in texture or mineral composition within the monzonitic and syenitic veins and schlieren; that is, none of these rock types show any chilling phenomena between each other.

The monzonite and syenite are sometimes coarse-grained, pegmatitic and rich in miarolitic cavities, suggesting abundant volatile components.

From these field observations the following conjectures can be drawn:

1. The monzonitic and syenitic parts are not isolated from the dolerite as independent bodies but are always confined within the doleritic parts in time as well as in space. Therefore, these intrusions are "composite intrusions" as defined by Thomas and Bailey (1924, p. 23).

2. The variation within the single intrusive body is a result of differentiation *in situ*.

MINERALS

Method of Investigation

In order to separate minerals from their parent rocks, finely crushed fragments of rocks were divided at first into felsic and mafic fractions by the use of Thoulet's solution of specific gravity about 3.0, and then pure samples were picked out by hand under the binocular microscope. Differences in color, cleavage, and shape made this separation method fairly accurate, but it was very tedious work. Stirring the heavy mineral assemblage in water greatly helped the separation of biotite flakes from amphibole and pyroxene, owing to the suspension of the flakes in the water.

Refractive indices were measured on the cleaved flakes by the usual immersion method, and optic angles and optic orientations were determined in thin section on a universal stage.

Chemical analyses of minerals as well as of rocks were made following the methods of Washington (1930) with some minor modifications. Total iron was determined by reduction of ferric oxide by zinc amalgam, and titration with 0.1N $KMnO_4$ solution. Titanium oxide is also reduced by this method, and a correction is necessary. Lime was sometimes determined by titration of calcium oxalate with $KMnO_4$ solution. The method is as accurate as the gravimetric one, and much quicker.

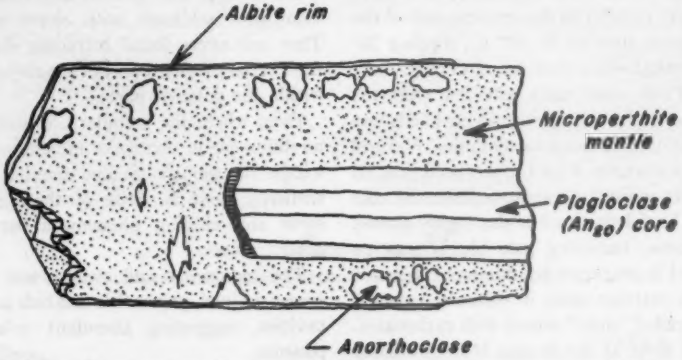

FIGURE 7. ZONAL STRUCTURE OF FELDSPAR No. 1219 IN MONZONITE FROM THE MOROTU RIVER
Length of the crystal is 15 mm.

Essential Minerals

FELDSPAR: Plagioclase plays a principal role in the dolerites, with subordinate anorthoclase either as a thin mantle or as an irregular mesostasis. Plagioclase ranges from labradorite to andesine. In the monzonites, feldspar always shows characteristic zoning composed of plagioclase, anorthoclase, or microperthite, and albite, passing from the core toward the margin. Plagioclase in the core varies from andesine to oligoclase. A thick shell of anorthoclase or microperthite constitutes the bulk of the crystal, while nearly pure albite forms the outermost thin mantle. In the syenites, plagioclase is subordinate or absent, and coarsely intergrown microperthite with thin albite stringers is the only feldspar. Anorthoclase is also absent. Descriptions are given of four analyzed feldspars.

Feldspar No. 1219.—Crystals of feldspar are developed in the miarolitic druses of the monzonite No. 1219 in the upper course of the Morotu River. One crystal studied is tabular parallel to (010), 16 by 14 by 6 mm. in size. Forms (010), (001), and (110) are observed, and cleavage parallel to (001) is perfect; tiny crystals of aegirinaugite are grown on this surface. The crystals show an interesting zonal structure with a core of plagioclase, a shell of microperthite or anorthoclase, and a rim of albite. This structure can be seen on the cleavage plane with the naked eye (Fig. 7).

Plagioclase in the core is twinned on the albite law, and $(010)\Lambda X' = 4°$, indicating An_{20}. It is homogeneous and fairly fresh, though secondary analcite is sporadically present. The shell, which forms a greater part of the crystal, is very fine microperthite with sporadic patches of anorthoclase and is twinned on the Carlsbad law. The boundary between plagioclase and anorthoclase or microperthite is quite sharp. Anorthoclase has $\alpha' = 1.524$, $\gamma' = 1.528$ on (001). The microperthite intergrowth coarsens toward the margin. The outermost thin rim, 0.01–0.1 mm. in width, is clear albite, twinned on the albite law. Sometimes this albite is continuous with the albite lamellae within the microperthite. Optical properties of albite are: $(010)\Lambda X' = 15° - 16°$, $\alpha' = 1.530$, $\gamma' = 1.539$ on (001), indicating An_{3-4}.

The chemical composition of the whole crystal is shown in column 1 of Table 2. Rather high Fe_2O_3 and MgO are ascribed to the presence of tiny aegirinaugite crystals, which were not eliminated because the whole crystal of feldspar was analyzed. The ratio of plagioclase to alkali feldspar shell is determined to be about 1:8 in volume, neglecting the thin albite rim, which is very subordinate in amount. If the composition of the plagioclase is taken as $Or_3\ Ab_{77}\ An_{20}$ (weight per cent), alkali feldspar has the calculated composition $Or_{23}\ Ab_{68}\ An_7$. It has a high content of Or molecule and can be distinguished from the other alkali feldspars which are rather rich in Ab molecule.

Feldspar No. 902.—The feldspar separated from the dolerite No. 902, at the Takara Bridge, consists of long prismatic crystals, 2 to 5 mm. long. It shows the same type of zoning as No. 1219. Plagioclase

TABLE 2.—FELDSPAR

Analyst: K. Yagi

	(1) 1219	(2) 902	(3) 8203	(4) 1412
SiO_2	64.41	62.14	63.84	66.39
TiO_2	0.10	0.09	0.14	0.07
Al_2O_3	19.35	21.58	19.08	19.39
Fe_2O_3	1.01	0.78	0.73	0.66
FeO	n.d.	n.d.	n.d.	n.d.
MnO	0.01	tr.	tr.	0.01
MgO	0.18	0.07	0.05	0.06
CaO	1.57	2.26	0.66	0.74
Na_2O	4.65	8.42	7.55	7.27
K_2O	7.84	3.18	4.77	5.53
H_2O+	n.d.	n.d.	n.d.	n.d.
H_2O-	0.57	1.59	2.34	0.33
Total.....	99.69	100.11	99.16	100.45
Bulk Feldspar Composition Wt. %				
Or	49.5	18.6	29.7	33.6
Ab	42.1	70.4	66.8	62.7
An	8.4	11.0	3.5	3.7
Composition of Alkali Feldspars Wt. %				
Or	55.3	22.5	32.0	36.5
Ab	37.8	71.0	66.0	61.0
An	6.9	6.5	2.0	2.5

1219: From monzonite, No. 1219, Morotu River
902: From monzonite, No. 902, Takara Bridge, Tiyo Mine
8203: From syenite, No. 8203, middle reaches of Morotu River
1412: From syenite, No. 1412, headwaters of Morotu River

in the core is fresh and twinned on the albite or Carlsbad law. Zonal structure is poorly developed, (010) $\Delta X' = 13°$–$16°$ indicating An_{30}. The microperthite shell is sometimes turbid because of sericitization and alteration to analcite. The boundary between the plagioclase and the microperthite is not so regular as in No. 1219, and no patches of homogeneous anorthoclase are found in the microperthite. Albite forms the outer rim around the microperthite. Chemical analysis (Table 2, column 2) shows a high soda content and rather low potash content. When feldspar molecules are calculated, there is an appreciable deficit in silica. This is probably due to secondary analcite along the cleavage which could not be completely removed. From the ratio of plagioclase and alkali feldspar, the chemical composition of the microperthite is calculated as $Or_{33}Ab_{71}An_6$. Compared to the alkali feldspar in No. 1219, the microperthite of this feldspar is very rich in the Ab molecule.

Feldspar No. 1412.—Crystals of feldspar, up to 2 cm. in length, are common in the drusy cavities of pegmatitic syenite No. 1412 in the ravine south of Mt. South Morotu. They are slightly bluish, tabular parallel to (010). Microscopic examination shows it is composed of microperthite with a very subordinate amount of plagioclase. The host orthoclase of the microperthite is slightly turbid, while the albite blebs, which sometimes form the outer rim of the microperthite, are clear. Alteration to analcite is rare. Albite $\alpha' = 1.529$, $\gamma' = 1.539$ on (001). Plagioclase is prismatic, twinned on the albite law, (010) $\Delta X' = 3°$, indicating An_{17}. The composition of the microperthite is calculated as $Or_{77}Ab_{21}An_3$, which is again rich in the Ab molecule (Table 2, column 4).

Feldspar No. 8203.—This feldspar in the syenite No. 8203 from the Morotu River is thick, tabular, and 0.5 to 5 mm. in size. It is composed chiefly of microperthite with the albite rim and a very subordinate amount of plagioclase core. The microscopic features are similar to those of No. 1412. Albite has $2V$ (+) = 74°, and the plagioclase is oligoclase, An_{20}. The composition of the microperthite is estimated as $Or_{32}Ab_{66}An_2$, which is very close to that of the microperthite of No. 1412.

OLIVINE: Olivine is an essential mineral in nearly all dolerites, but is absent in the monzonites or syenites. Sometimes it forms phenocrysts about 1 mm. in size in the porphyritic dolerites. It is occasionally surrounded by the titanaugite or titanbiotite, but there is no definite relation between their crystallographic orientations. Olivine is usually altered to serpentine, with or without calcite. The serpentine shows various shades of green or yellow, with low birefringence. Because of the alteration, exact optical properties cannot be obtained; $2V$ (−) = 74°, 75°, 78°, 80°. From these values the range of composition of olivine is estimated as Fa_{37-50}, and it can be called hyalosiderite (Wager and Deer, 1939, p. 75).

PYROXENE: Pyroxenes are the most important mafic minerals throughout the whole series of rocks and play the chief role in the evolution of mafic minerals. They are monoclinic and their prevalent zonal structure shows a very wide range of composition. Zonal crystals, from soda-augite through aegirinaugite to almost pure aegirine, are common

in some monzonites or syenites. The pyroxenes are described in the order of their crystallization.

Diopsidic augite.—The earliest stage in the evolution of the pyroxenes is represented by augite in the pea-shaped aggregate of some dolerites. The optical properties of augite from the aggregate in the dolerite at Tyaki are: $\gamma' = 1.721$ on (010), $2V(+) = 60°$, $c\Lambda Z = 42.5°$–$45°$, pale brown or pale violet, nonpleochroic. Although this augite was not chemically analyzed, its composition is estimated from the optical properties as well as from the bulk composition of the aggregate as diopsidic augite, $Wo_{52}En_{39}Fs_9$ (weight per cent). Similar diopsidic augite is found also as phenocrysts in some porphyritic dolerites of the area.

Titanaugite.—Titanaguite is most prevalent in the dolerites and is also found in some monzonites. In some coarse-grained dolerites, as at Takara Bridge, Tiyo Mine, or at Morotu Cape, large crystals 5 to 10 cm. long and 1 to 2 cm. wide form radiating aggregates. However, usually the size is 1 to 3 mm. in the dolerites and smaller in the monzonites. Crystal forms observed are (100) (010) (110) ($\bar{1}$01) and ($\bar{1}$11). Some crystals are stout prismatic or tabular but more commonly they are subhedral and irregular showing typical ophitic structure with euhedral plagioclase crystals. Twinning on (100) is very common. Zonal structure in the titanaugite is not marked, but it is frequently surrounded by slightly greenish augite apparently rich in soda. Although hourglass structure is absent, some large crystals are composed of numerous irregular parts slightly different in their optical orientation and properties. Parallel intergrowth with other mafic minerals is very common.

Barkevikite is frequently intergrown with titanaugite, with the b and c axes in common; the augite is then usually surrounded by a thick layer of hornblende, but the reverse case is also found. Sometimes barkevikite is formed around the titaniferous iron ores enclosed in the titanaugite, suggesting that the barkevikite is produced by reaction between the two minerals. Some titanbiotite also shows peripheral growth on augite. Some olivine is found in the core of augite, but the two minerals do not show any crystallographic relation. Titanaugite, though more resistant to weathering or other secondary alteration than feldspars or olivine, is in some places completely replaced by carbonates and chlorites. Chemical analyses and optical properties of two titanaugite samples (Nos. 1507 and 607) separated from dolerites from Takara Bridge and Morotu Cape, respectively, are given in Table 3. These two samples are similar in composition and are characterized by high titanium.

Opinions differ concerning the formula of titanaugite. Machatschki (1929, p. 228) and Warren and Biscoe (1931, p. 399) assumed that, because of the similarity of the ionic radii, Ti replaces Mg with the formula $X_mY_{2-m}(Si, Al)_2(O, OH, F)_6$ where X = Ca, Na, K, Y = Fe″, Fe‴, Mn, Mg, Ti. Berman (1937, p. 354) and Hess (1949, p. 624), taking electrical charge into consideration, showed that $2Al^{+3}$ should replace $2Si^{+4}$ if Ti^{+4} replaces Mg^{+2}. Kunitz (1930, p. 240), however, admitted the isomorphous replacement of Si by Ti diadochically. This problem has been also discussed by Barth (1931, p. 223) who concluded from his experimental study that Ti can replace Si as well as Mg in the pyroxene structure. Despite differences in ionic radii, Dixon and Kennedy (1933, p. 115) and

TABLE 3.—TITANAUGITE

Analyst: K. Yagi

	Wt. per cent			Atomic ratio	
	1507	607		1507	607
SiO_2	48.99	46.76	Si	183	177
TiO_2	2.83	2.96	Ti	8	8
Al_2O_3	3.29	6.26	Al	14	28
Fe_2O_3	2.16	1.45	Fe‴	6	4
FeO	7.56	8.92	Fe″	24	28
MnO	0.05	0.16	Mn	0	0
MgO	13.24	12.07	Mg	74	69
CaO	21.72	20.34	Ca	87	82
Na_2O	0.49	0.43	Na	4	3
K_2O	0.22	0.28	K	1	1
H_2O+	n.d.	n.d.			
H_2O-	0.22	0.89	O	600	600
Total.....	100.77	100.52			

	No. 1507	No. 607
Pleochroism		
X	pale reddish brown	brownish purple
Y	pale brown	brownish purple
Z	pale greenish brown	pale brown
	X > Y > Z	X > Y > Z
2V(+)	33.5°, 37°, 41°	38°, 50°, 53°
c Λ Z	36°–43°	36°–43°
α′	1.703 on (110)	1.694 on (110)
γ′	1.723 on (110)	1.719 on (110)

1507: From dolerite, No. 1507, Takara Bridge, Tiyo Mine

607: From dolerite, No. 607, Morotu Cape

Deer and Wager (1938, p. 17) argue that Si may be replaced by Ti when combined Si and Al are insufficient to satisfy the pyroxene chain. Rankama and Sahama (1950, p. 561) also admitted the limited replacement of Si by Ti. In the titanaugite No. 1507 and titaniferous soda-augite No. 1305G there is apparent deficiency in Si + Al. The writer follows Deer and Wager in assuming that Ti takes the position of Si. For the consistency of the argument, this method will be followed even in those cases where there is no deficiency in Si + Al.

The formulas of the titanaugites follow:

Titanaugite No. 1507: $(Ca, Na, K)_{0.9}(Fe'', Fe''', Mn, Mg, Al)_{1.1}(Si, Ti, Al)_2O_6$

Titanaugite No. 607: $(Ca, Na, K)_{0.9}(Fe'', Fe''', Mn, Mg, Al)_{1.1}(Si, Ti, Al)_2O_6$

In both cases the sum of X group atoms (Ca, Na, K) is smaller than unity, and that of Y group (Fe'', Fe''', Mn, Mg, Al) is larger than unity, and the total of the two groups is nearly twice unity. Therefore these formulas are in close agreement with the generally accepted pyroxene formula,

$$X_mY_{2-m}(Si, Al)_2(O, OH, F)_6.$$

Soda-augite.—Soda-augite is most prevalent in the monzonites and occurs either as isolated crystals or as the core of aegirinaugite. It also forms the outer rim on the titanaugite in the dolerites. The size ranges from 0.2 to 1.0 mm. and rarely attains 5 mm. Ophitic texture, so common in titanaugite, is usually absent. It is light green to almost colorless and pleochroism is generally weak. Some titaniferous soda-augite shows a violet tint. In the zonal structure, outer zones are always a deeper green than the core. Coronal growth with other mafic minerals is very common. In syenites, hastingsitic hornblende and soda-augite are in parallel growth, both having the b and c axes in common. Sometimes soda-augite grows on a kaersutite crystal, both being further mantled by titanbiotite. Titaniferous soda-augite in the monzonite No. 1305G at Tyaki, Tiyo, and soda-augite in the monzonite No. 902 at Takara Bridge, Tiyo Mine, were separated from their parent rocks and chemically analyzed (Table 4). Hourglass structure is well developed, the cones on the prismatic zone (100) are deeper in color than those on (111), and the pleochroism is also different.

The two samples of soda-augite are similar in chemical composition, although titanium is high in augite No. 1305G, which is intermediate between titanaugite and soda-augite. In both samples, the sum of Si + Al falls short of 200, and addition of Ti to this group is justified, admitting the isomorphous replacement of Si — Ti. The chemical formulas follow:

Titaniferous soda-augite No. 1305G: $(Ca, Na, K)_{0.9}(Fe'', Fe''', Mn, Mg)_{1.1}(Si, Ti, Al)_2O_6$

Soda-augite No. 902: $(Ca, Na, K)_{0.9}(Fe'', Fe''', Mn, Mg)_{1.2}(Si, Ti, Al)_2O_6$

TABLE 4.—SODA-AUGITE

Analyst: K. Yagi

	Wt. per cent			Atomic ratio	
	1305G	902		1305G	902
SiO_2	46.32	48.30	Si	177	178
TiO_2	2.76	1.42	Ti	8	4
Al_2O_3	2.64	3.39	Al	12	15
Fe_2O_3	6.30	3.82	Fe'''	18	11
FeO	8.93	11.69	Fe''	28	36
MnO	0.28	0.17	Mn	1	0
MgO	11.91	10.95	Mg	68	61
CaO	19.53	18.28	Ca	80	72
Na_2O	1.69	1.62	Na	12	12
K_2O	0.26	0.29	K	2	1
H_2O+	n.d.	n.d.			
H_2O-	0.20	0.55	O	600	600
Total.....	100.82	100.48			

	No. 1305G	No. 902
Pleochroism		
X	pale greenish violet	pale green or grass green
Y	pale violet	pale green or grass green
Z	pale brownish violet	pale yellowish green or yellowish green
	X > Y ≥ Z	X > Y ≥ Z
2V(+)	58°, 62°, 64°, 72°	54°, 66°
c Δ Z	45.5°, 49.5°	42°
α'	1.694 on (110)	1.698 on (110)
γ'	1.719 on (110)	1.721 on (110)

1305G: From monzonite, No. 1305G, Tyaki, Tiyo Village

902: From monzonite, No. 902, Takara Bridge, Tiyo Mine

Aegirinaugite.—Aegirinaugite occurs as isolated crystals in some of the monzonites or syenites, but more frequently it shows remarkable zonal growth with soda-augite and aegirine. Euhedral prismatic crystals are found in some cases, but its crystalliza-

tion in the syenites is apparently later than that of feldspar, which sometimes assumes euhedral outlines toward the aegirinaugite. Twinning on (100) is common, and cleavage parallel to (110) is perfect. It is deep green with strong pleochroism, the margin always being deeper in color than the core of the zoned crystals. Parallel growth with hastingsite, arfvedsonite, aenigmatite, and aegirine is common. Although the aegirinaugite is usually fresh, there is some alteration to chlorite with high birefringence. In syenite some aegirinaugite crystals are substantially replaced by a brownish-yellow chloritic mineral with the following pleochroism: X' = golden yellow, Z' = golden yellow with brownish tinge. Upon further alteration this mineral becomes colorless with very faint birefringence, resembling opal. It is determined as nontronite.

TABLE 5.—AEGIRINAUGITE
Analyst: K. Yagi

	Wt. per cent			Atomic ratio	
	8203	1412K		8203	1412K
SiO_2	44.26	47.82	Si	177	189
TiO_2	2.38	1.19	Ti	7	4
Al_2O_3	4.76	2.64	Al	23	12
Fe_2O_3	15.48	22.36	Fe‴	47	66
FeO	11.53	10.58	Fe″	38	35
MnO	0.70	0.00	Mn	2	0
MgO	4.10	0.94	Mg	25	6
CaO	10.64	6.14	Ca	46	26
Na_2O	3.76	7.78	Na	29	59
K_2O	0.85	0.16	K	4	1
H_2O+	n.d.	n.d.			
H_2O-	2.12	1.16	O	600	600
Total.....	100.58	100.77			

	No. 8203	No. 1412K
Pleochroism		
X	pale bluish green	bluish green
Y	pale green	green
Z	yellowish green	yellowish green
	X > Y > Z	X > Y > Z
2V(+)	53°, 56°, 65°, 72°	74°, 75.5°
c ∧ Z	60°	66°
α'	1.708 on (110)	1.721 on (110)
γ'	1.738 on (110)	n.d. on (110)

8203: Aegirinaugite crystal from syenite, No. 8203, middle reaches of Morotu River
1412K: Aegirinaugite core of aegirine crystal from syenite, No. 1412, headwaters of Morotu River

The chemical composition and optical properties of aegirinaugite No. 8203 from the syenite at the Morotu River are given in Table 5. Large crystals of aegirine in the pegmatitic syenite No. 1412 from the headwaters of the Morotu River show strong zoning with a core of aegirinaugite, a thick deep green shell of aegirine, and a pale outer rim of pure aegirine. Owing to the remarkable difference in color as well as the shape of cleaved flakes, it was possible to separate the different zones in the crushed powder under the binocular microscope, and a chemical analysis was made of the aegirinaugite core (1412K, Table 5), of the pure aegirine rim (1412H,

TABLE 6.—AEGIRINE
Analyst: K. Yagi

	Wt. per cent				Atomic ratio		
	1411	1412S	1412H		1411	1412S	1412H
SiO_2	49.06	51.30	51.51	Si	191	199	198
TiO_2	2.34	0.72	0.68	Ti	7	2	2
Al_2O_3	2.36	3.08	2.86	Al	11	14	13
Fe_2O_3	22.69	24.32	28.12	Fe‴	66	71	81
FeO	7.45	6.18	3.01	Fe″	24	20	10
MnO	0.68	0.08	0.00	Mn	2	0	0
MgO	1.48	0.57	0.04	Mg	9	3	0
CaO	4.80	3.08	1.80	Ca	20	13	7
Na_2O	8.25	9.25	11.48	Na	62	69	85
K_2O	0.52	0.18	0.32	K	3	1	2
H_2O+	n.d.	n.d.	n.d.				
H_2O-	0.71	0.78	0.24	O	600	600	600
Total...	100.34	99.54	100.06				

	No. 1411	No. 1412H
Pleochroism		
X	bluish green	pale straw yellow
Y	green	colorless
Z	greenish yellow	colorless
	X > Y > Z	X > Y = Z
2V(−)	66.5°	60°
c ∧ Z	85°	89°
α'	1.748–1.771 on (110)	1.748–1.774 on (110)
γ'	n.d.	n.d.

1411: Aegirine crystal from syenite No. 1411, headwaters of Morotu River
1412S: Aegirine crystal as a whole from syenite, No. 1412, headwaters of Morotu River
1412H: Aegirine rim of aegirine crystal from syenite, No. 1412, headwaters of Morotu River

Table 6), and also of aegirine crystal as a whole (1412S, Table 6). The chemical composition and optic properties of the aegirinaugite No. 1412K are also given in Table 5. The aegirinaugite No. 1412K is much richer in Na_2O and Fe_2O_3, while poorer in CaO and MgO, than the aegirinaugite No. 8203 in which the rather high K_2O replaces a considerable portion of Na_2O. The chemical formulas of the aegirinaugites follow:

Aegirinaugite No. 1412K: $(Na, K, Ca)_{0.9}(Fe''', Fe'', Mg, Al)_{1.1}(Si, Ti, Al)_2O_6$

Aegirinaugite No. 8203: $(Na, K, Ca)_{0.8}(Fe''', Fe'', Mn, Mg, Al)_{1.2}(Si, Ti, Al)_2O_6$

Aegirine.—Aegirine is most abundant in syenites. In some pegmatitic facies crystals as large as 10 mm. are not uncommon. They are bounded by the forms (100), (010), (110), (101), and (1̄11). In some cases aegirine is older than the feldspars, but usually aegirine forms anhedral crystals in the mesostasis between feldspars, suggesting that its formation was, for the most part, later than the formation of albite layers around the feldspars.

Besides these large crystals, aggregates of slender prismatic crystals or fine hairlike crystals radiating from a common center are found in the small drusy cavities. They are elongated along the c axis and attain 0.05 to 0.1 mm. in length. These small crystals belong to the last stage of crystallization and perhaps are the product of hydrothermal processes.

Large crystals are bluish green, grass green, or pale green. If they are zoned, the outer zones are always paler with increase of the aegirine molecule. The outermost zone is nearly pure aegirine and is very pale green or almost colorless. In the zoned crystals, therefore, the color deepens as it extends from the core until it attains maximum intensity, when it again diminishes to an almost colorless rim. Some thin needle crystals of aegirine are also nearly colorless. Parallel growth with arfvedsonite, aenigmatite, etc., is observed; small hematite flakes are sometimes included in the aegirine, showing the reaction relation between the two minerals found in the laboratory (Bowen, Schairer, Willems, 1930, p. 431). Though aegirine is generally fresh, some has been altered to a reddish-brown substance, which is probably iron hydroxide, and a yellow nontronitic mineral with high birefringence. Cleavage parallel to (110) is perfect. The chemical composition and optical properties of aegirines in the syenites at the Morotu River are given in Table 6. No. 1411 is an analysis of complete crystals showing slight zoning; No. 1412S is the bulk analysis of strongly zoned crystals; and No. 1412H is the pure aegirine rim of the same crystals. Optical properties also show wide variation.

Chemical formulas follow:

Aegirine No. 1411: $(Na, K, Ca)_{0.9}(Fe''', Fe'', Mn, Mg, Al)_{1.1}(Si, Ti, Al)_2O_6$

Aegirine No. 1412S: $(Na, K, Ca)_{0.8}(Fe''', Fe'', Mg, Al)_{1.1}(Si, Ti, Al)_2O_6$

Aegirine No. 1412H: $(Na, K, Ca)_{0.9}(Fe''', Fe'', Mg, Al)_{1.0}(Si, Ti)_2O_6$

Samples No. 1411 and No. 1412S are much alike; Sample No. 1412H is a pure aegirine richest in acmite molecule.

Amphiboles: Amphiboles in these rocks show wide variation. In the dolerites brown barkevikitic hornblende is intimately associated with titanaugite or titanbiotite. Kaersutite occurs abundantly in some monzonites and syenites. In these rocks kaersutite rarely is surrounded by a mantle of hastingsitic hornblende. Arfvedsonite is the only amphibole in most of the alkalic syenites, where it frequently occurs in parallel growth with aegirine. Aenigmatite, which is sometimes grouped in the amphibole family, occurs in some places in the monzonites and syenites. Common hornblende and rhombic amphiboles are absent. Only kaersutite and arfvedsonite were chemically analyzed; the other amphiboles were determined optically.

Barkevikite.—Brown hornblende is included in titanaugite or titanbiotite, or less often as isolated small crystals in some dolerites from Morotu Cape. In these rocks iron ore inclusions in titanaugite are often surrounded by a thin rim of hornblende, 0.01 to 0.1 mm. wide, which always has crystallographic orientation in common with the host augite crystals. Pleochroism is marked, but less intense than that of kaersutite.

X = pale violet brown Y = violet brown
Z = reddish brown $c\Lambda Z = 2.5°$
$2V(-) = 80°$ (core) $\rightarrow 77°$ (margin)

By virtue of the above properties, this amphibole is clearly distinguished from common hornblende, but it is less titaniferous than kaersutite and is therefore designated barkevikite. As the two amphiboles do not occur in the same rock, the genetic relation of barkevikite and kaersutite is not certain.

Kaersutite.—Dark brown or almost black lustrous crystals of hornblende can be observed in the monzonite veins traversing the dolerite at Tyaki, along the Tiyo River. Chemical analysis of this hornblende shows unusually high titanium, by which the hornblende is determined as kaersutite, a rather rare variety of barkevikite. This mineral also occurs in the syenite from the triangulation point north to Tyaki.

Prismatic crystals, usually 0.5 to 3.0 mm. (and, rarely, 10 mm.) long, are bounded by the forms

(100), (010), with the basal pinacoid less prominent. Cleavage parallel to (110) is perfect, and twinning on (100) is common. Besides common inclusions, such as apatite, magnetite, or ilmenite, the crystals sometimes include featherlike light gray inclusions arranged diagonally in parallel rows. These inclusions have definite orientation relative to the cleavage of the host kaersutite crystals, but their extremely minute size makes identification impossible.

Parallel growth with other mafic minerals is worthy of mention. The order of arrangement from the core to the margin is as follows:

Kaersutite → titanbiotite → sodic titanaugite
Kaersutite → sodic titanaugite → titanbiotite

In these cases, b and c axes are in common. The boundary between them is sharp and irregular. Magmatic resorption or an opacite margin, so common in the basaltic hornblende of the volcanic rocks, is absent.

TABLE 7.—KAERSUTITE No. 1305G

Analyst: K. Yagi

	Wt. per cent		Atomic ratio
SiO_2	40.73	Si	604
TiO_2	8.47	Ti	94
Al_2O_3	10.62	Al	185
Fe_2O_3	1.39	Fe‴	16
FeO	11.91	Fe″	147
MnO	0.25	Mn	4
MgO	10.46	Mg	233
CaO	10.40	Ca	166
Na_2O	2.92	Na	84
K_2O	1.33	K	25
H_2O+	1.97	OH	194
H_2O-	0.23	O	2206
Total.....	100.68		

Pleochroism	
X	brownish yellow or straw yellow
Y	reddish brown
Z	deep chestnut brown with violet tinge
Absorption	Z > Y > X
Extinction angle	
c ∧ Z	16°, 17°, 19°
Optic angle	
2V(−)	77°, 78°, 79°, (in the core) → 73° (in the margin)
Index	
α'	1.680 on (110)
γ'	1.703 on (110)

The chemical composition and optical properties of the kaersutite from the monzonite No. 1305G at Tyaki are given in Table 7. During the chemical analysis the usual fusion by alkali carbonates failed to render all the material completely soluble in HCl, and therefore the fusion was repeated. Impurities in the silica remaining after the treatment by HF were unusually high and probably were largely TiO_2.

Kaersutite is a rare mineral, and there are so far only six known localities of analyzed kaersutite. They are Kaersut, Greenland—original locality (Lorenzen, 1886, p. 318; Gossner and Spielberger, 1930, p. 120; Drescher, 1933, p. 247); Linosa, Italy (Washington and Wright, 1908, p. 1870); Tikaisi, Oki Island, Japan (Tomita, 1934, p. 99); Uturyoto, Korea (Harumoto, 1933, p. 96); Yohodo, Korea (Kawano, 1934, p. 124); and Boulder Dam, U. S. A. (Campbell and Schenk, 1950, p. 684). When the chemical analyses of these are compared (Table 8), it is apparent that kaersutites can be classified into two groups—the one characterised by high Fe_2O_3 and low FeO, and the other by low Fe_2O_3 and high FeO, while the sum of iron oxides is rather constant throughout both groups. Kaersutite from Linosa, Yohodo, Uturyoto, and Boulder Dam belong to the first group, and they always occur in the basaltic rocks. Kaersutite from Kaersut and Morotu fall in the second group and occur in the leucocratic veins traversing the mafic rocks. Therefore, the difference in their chemical compositions might be ascribed to the difference in the physico-chemical environment of their parent rocks.

In the calculation of the chemical formula, the sum of Si + Al in the atom group is less than 8, when we take O + OH = 24, and therefore there should be replacement of Si-Ti in this kaersutite. Thus it is expressed as follows: $OH_{1.9}(Ca, Na, K)_{2.2}(Fe'', Fe''', Mn, Mg, Al)_{4.9}(Si, Ti, Al)_8O_{22}$. This is in agreement with Warren's general hornblende formula of $(OH, F)_2(Ca, Na, K, Mn)_{2-3}(Mg, Fe, Ti, Mn, Al)_5(Si, Al)_8O_{22}$ (Warren, 1930, p. 516).

Hastingsite.—Brownish-green hornblende is an essential mineral in some monzonite from the Morotu River. It is generally stout prismatic and occurs sometimes as isolated crystals, but usually in parallel growth with other mafic minerals. In some cases, soda-augite is surrounded by a thick mantle of this hornblende with the b and c axes in common, and the boundary is sharp. In other cases, kaersutite has narrow, incomplete titanbiotite fringes which are succeeded by hastingsite, the latter often coming into direct contact with the

kaersutite and forming parallel growth with the b and c axes in common. Zonal structure in hastingsite itself is sometimes observed. The color becomes surrounded by arfvedsonite with gradual transition, occurs in some syenites and monzonites of this district.

TABLE 8.—KAERSUTITE

Analyst	Yagi	Teshima	Kawano	Usizima	Gossner	Washington	Gonyer
	(1)	(2)	(3)	(4)	(5)	(6)	(7)
SiO_2	40.73	39.78	38.30	39.20	39.50	40.85	41.46
TiO_2	8.47	7.00	6.02	6.53	10.33	8.47	5.70
Al_2O_3	10.62	14.13	12.87	13.87	11.12	9.89	14.24
Fe_2O_3	1.39	4.61	7.98	4.08	0.06	8.85	3.32
FeO	11.91	7.31	6.96	7.33	9.44	3.96	5.70
MnO	0.25	0.12	0.12	0.11	0.10	0.12	0.08
MgO	10.46	11.01	11.79	11.96	12.90	12.47	13.68
CaO	10.40	10.75	10.47	12.37	10.91	12.16	11.62
Na_2O	2.92	2.57	3.11	1.99	3.82	2.01	2.29
K_2O	1.33	1.58	1.30	1.45	1.43	0.63	1.72
H_2O+	1.97	0.37	1.10	0.87	0.59	0.19	0.12
H_2O-	0.23	0.22	0.26	0.27	—	—	—
F	—	—	0.05	—	—	0.28	0.42
	100.68	99.45	100.33	100.03	100.20	99.98*	100.35
Less 0 = F	—	—	0.02	—	—	0.12	0.18
Total.....	100.68	99.45	100.31	100.03	100.20	99.86	100.17

* Includes NiO, 0.10

1. From Tyaki, Morotu
2. From Tikaisi, Dogo, Oki, Japan (Tomita, 1934, p. 1)
3. From Yohodo, Korea (Kawano, 1934, p. 127)
4. From Uturyo Island, Korea (Harumoto, 1933, p. 100)
5. From Kaersut, Greenland (Gossner and Spielberger, 1930, p. 121)
6. From Monte Rosso, Linosa (Washington and Wright, 1908, p. 192)
7. From Boulder Dam, Arizona (Campbell and Schenk, 1950, p. 684)

more greenish, and both the extinction angle and the optic angle increase toward the margin. Optical properties follow:

Pleochroism:

X = pale brownish yellow—pale brownish green
Y = deep greenish brown—pale greenish brown
Z = deep greenish brown—bluish green

Absorption $Z > Y > X$

$2V(-) = 48°, 48.5°, 54°$ (core) $\rightarrow 64°$ (margin)
$c\wedge Z = 5°$ (core) $\rightarrow 10° \rightarrow 20°$ (margin)

From these properties the hornblende should probably be regarded as hastingsite. The mode of zoning with other amphiboles shows that hastingsite is preceded by kaersutite, while it is earlier than arfvedsonite. Iwao (1939, p. 180) has mentioned that quasi-uniaxial hastingsite, which is always

Arfvedsonite.—Arfvedsonite is confined to alkali-rich monzonites or syenites with abundant aegirine. It occurs either in isolated crystals or in parallel growth with other mafic minerals. It is broad tabular or prismatic, deep bluish green with strong pleochroism. The size ranges between 0.02 mm. and 4 mm. but is usually 0.1 to 0.5 mm. Arfvedsonite sometimes occurs intergrown with aegirinaugite, aegirine, and hastingsite. Crystals of aegirine and arfvedsonite in parallel growth in the syenite No. 1412 from the headwaters of the Morotu River were separated from the parent rock. They were then crushed, and the arfvedsonite sample was carefully picked out under the binocular microscope. The chemical composition and optical properties of this arfvedsonite are given in Table 9.

Arfvedsonite in these alkalic rocks does not give perfect extinction on (010) plane between the crossed nicols, even in monochromatic light. Eskola and

Sahlstein (1930, p. 94) explained this phenomenon, occasionally observed in alkali hornblende, as due to the very fine perthitic exsolution lamellae. Sato (1936, p. 141) suggested that the optical abnormality

TABLE 9.—ARFVEDSONITE No. 1412
Analyst: K. Yagi

	Wt. per cent		Atomic ratio	
SiO_2	48.51	Si	756	
TiO_2	1.10	Ti	13	790
Al_2O_3	1.08	Al	21	
Fe_2O_3	17.53	Fe‴	204	
FeO	14.31	Fe″	186	429
MnO	0.66	Mn	8	
MgO	1.31	Mg	31	
CaO	5.43	Ca	91	
Na_2O	7.87	Na	237	358
K_2O	1.48	K	30	
H_2O+	1.48	OH	153	2400
H_2O-	0.17	O	2247	
Total.....	100.93			

Pleochroism	
X	greenish brown
Y	greenish blue
Z	deep greenish blue with violet tinge
Absorption	
Z > Y > X	
Extinction angle	
c ∧ X	5°–6°
Index	
α'	1.686 on (110)
γ'	1.695 on (110)

of alkali amphibole in granite from Jehol, Manchuria, was presumably a result of its optical pseudo-activity. Iwao (1939, p. 183) remarked that the two pseudo-vibration directions of polarized light perpendicular to (010) do not intersect at right angles. This phenomenon was later discussed by Hori (1942) who concluded that the particles of the alkali hornblendes are arranged in random position in the (010) plane.

The analysis (Table 9) shows that CaO content is high, and total iron is also high, while MgO and Al_2O_3 are low. On the basis of OH + O = 24, the following formula is obtained: $OH_{1.5}(Na, K, Ca)_{2.6}$ (Fe″, Fe‴, Mn, $Mg)_{4.2}(Si, Ti, Al)_{7.9}O_{22.5}$. This is somewhat different from the accepted hornblende formula of $OH_2(Na, K, Ca)_{2.3}$ (Fe″, Fe‴, Mg, Ti, Mn, $Al)_5(Si, Al)_8O_{22}$. The group (Na, K, Ca) is too high, and the group (Fe″, Fe‴, Mg, Mn) is too low. However, if we accept the isomorphous replacement of Ca-Mg in this amphibole, just as in the pyroxene, we get the formula, $OH_{1.5}(Na, K)_{2.7}$ (Fe″, Fe‴, Mn, Mg, $Ca)_{5.2}$ (Si, Ti, $Al)_{7.9}O_{22.5}$, which is in close agreement with the general hornblende formula.

Aenigmatite.—The crystallographic position of aenigmatite has not yet been settled, and it is sometimes omitted from the amphibole group (Fleischer, 1936, p. 346). In the present discussion, however, it is tentatively included.

Aenigmatite is a rather rare mineral and in this area is found only in some monzonites and syenites from the Morotu River. It seldom attains 1 mm. in size and is usually less than 0.1 mm. Cleavage parallel to (110) is well developed, and twinning is not observed. Pleochroism is extraordinarily strong and characteristic.

X = reddish brown
Y = deep chestnut brown
Z = dark brown or nearly opaque
Extinction angle $c \wedge Z = 27°–55°$
$\alpha' = 1.774$ on (110)

In the large crystals zonal structure is present; the core is deep greenish brown with the following strong pleochroism:

X = slightly greenish brown
Y = deep brown
Z = very dark brown to opaque

This pleochroism differs from that of aenigmatite and suggests that the mineral in the core is rhönite, though the exact optical properties could not be determined. Optical orientation is the same throughout the zoned crystals, but the boundary between the two minerals is very sharp. Therefore, it seems probable that rhönite and aenigmatite form a discontinuous reaction series, of which the earlier member is rhönite (Fleischer, 1936, p. 348). A similar relation was observed by Tomita (1934, p. 124) in the resorption products of kaersutite from Titaisi, Oki, Japan. It was not possible to obtain sufficient material for chemical analysis.

BIOTITE: Biotite is an important mafic mineral in all dolerites of the area, and in some monzonites also; but in syenites it is either subordinate in amount or absent. Biotite in the dolerites and monzonites is characterized by a deep reddish-brown color and strong pleochroism due to high content of titanium and should be called titanbiotite. The biotite in the syenites is a lepidomelanic variety characteristically deep brown with a greenish tinge.

Titanbiotite.—In dolerites titanbiotite forms small individual crystals or large crystals ophitically related to plagioclase. It is usually fresh, but may be altered to chlorite. The chemical composition and

TABLE 10.—TITANBIOTITE

Analyst: K. Yagi

	Wt. per cent			Atomic ratio			
	607	1305G		607		1305G	
SiO_2	33.60	36.12	Si	260		270	
TiO_2	5.79	8.27	Ti	33	425	46	424
Al_2O_3	14.47	12.18	Al	132		108	
Fe_2O_3	7.00	2.21	Fe‴	41		13	
FeO	11.20	15.40	Fe″	73	250	96	245
MnO	0.30	tr.	Mn	2		0	
MgO	11.56	12.15	Mg	134		136	
CaO	2.39	1.82	Ca	20		17	
Na_2O	0.41	1.15	Na	7	109	17	95
K_2O	8.29	6.40	K	82		61	
H_2O+	3.12	4.03	OH	161	1200	200	1200
H_2O-	2.00	0.96	O	1039		1000	
Total...	100.13	100.69					

	No. 607	No. 1305G
Pleochroism		
X	pale straw yellow	pale yellow or pale brownish yellow
Y = Z	dark reddish chestnut brown	dark reddish brown
	Y = Z > X	Y = Z > X
Optic angle		
2V(−)	0°, rarely 3°	0°, rarely 16°
Index		
γ	1.659	1.657

607: From dolerite, No. 607, Morotu Cape
1305G: From monzonite, No. 1305G, Tyaki, Tiyo Village

optical properties of titanbiotites in the dolerite (No. 607) at Morotu Cape and in the monzonite vein (No. 1305G) at Tyaki are given in Table 10. High titanium in both biotites is characteristic. The high CaO in the biotite No. 607 is partly due to impurity of titanaugite which could not be completely eliminated from the analyzed sample. However, the analyzed sample of biotite No. 1305G was very pure, and the CaO determined must be regarded as an ingredient of the biotite. Tsuboi, Sugi, and Iwao (1938, p. 453) have remarked that biotites of igneous origin have more or less CaO, while those in the metamorphic rocks have a very subordinate amount. The content of CaO in this biotite is the highest even among the biotites of igneous origin. Na_2O is also rather high, K_2O is accordingly low, and Na as well as Ca takes the place of K. Nockolds (1947, p. 401) discussed the relation between chemical composition and paragenesis in biotites of igneous origin. When the two biotites are plotted in a triangular diagram of MgO-FeO[1]-Al_2O_3, they fall within the field of biotites associated with hornblende, pyroxene, or olivine. This agrees with the actual paragenetic relation in these rocks. It is also noticed that Al_2O_3 content is low in these biotites, especially in the biotite No. 1305G, compared with biotites in the calcalkalic rocks.

The chemical formulas, computed on the basis of OH + O = 12, are:

Titanbiotite No. 1305G: $OH_2(K, Na, Ca)_{0.9}$ (Fe″, Fe‴, Mn, Mg, Al)$_{2.7}$(Si, Ti, Al)$_4O_{10}$

Titanbiotite No. 607: $(OH, O)_2(K, Na, Ca)_{1.1}$(Fe″, Fe‴, Mn, Mg, Al)$_{2.8}(Si, Ti, Al)_4O_{10}$

They are in close agreement with the general biotite formula, $(OH, F)_2(K, Na, Ca)_1(Fe'', Fe''', Mg)_{2-3}Si_4O_{10}$. In some biotite OH is less than 2, and the deficiency in OH is supplied by O. Prider (1939, p. 373) also has remarked that OH is usually lower than unity in the formula of titanium-rich biotites. In the present case, the titanbiotite No. 607 has a deficiency of OH, and the titanbiotite No. 1305G, which has higher titanium, has just enough OH to satisfy the formula.

Lepidomelanic biotite.—Titanbiotite is absent in nearly all the syenites, and its place is taken by small flakes of deep greenish-brown biotite, 0.05 to 0.3 mm. in size. Small radial aggregates of such biotite are found in some syenites, especially in close association with aegirine.

Pleochroism:

X = pale brownish yellow
Y = *Z* = dark chestnut brown with greenish tinge

Although no chemical analysis of this biotite was made, it is probably lepidomelanic biotite, rich in Fe_2O_3 and rather poor in TiO_2, judging from its characteristic color (Hall, 1941, p. 32). This inference is in accord with the bulk composition of the syenites.

IRON ORES: The abundance of iron ores and their remarkable association with other mafic minerals

[1] All Fe is expressed as FeO.

tion in the syenites is apparently later than that of feldspar, which sometimes assumes euhedral outlines toward the aegirinaugite. Twinning on (100) is common, and cleavage parallel to (110) is perfect. It is deep green with strong pleochroism, the margin always being deeper in color than the core of the zoned crystals. Parallel growth with hastingsite, arfvedsonite, aenigmatite, and aegirine is common. Although the aegirinaugite is usually fresh, there is some alteration to chlorite with high birefringence. In syenite some aegirinaugite crystals are substantially replaced by a brownish-yellow chloritic mineral with the following pleochroism: X' = golden yellow, Z' = golden yellow with brownish tinge. Upon further alteration this mineral becomes colorless with very faint birefringence, resembling opal. It is determined as nontronite.

TABLE 5.—AEGIRINAUGITE

Analyst: K. Yagi

	Wt. per cent			Atomic ratio	
	8203	1412K		8203	1412K
SiO_2	44.26	47.82	Si	177	189
TiO_2	2.38	1.19	Ti	7	4
Al_2O_3	4.76	2.64	Al	23	12
Fe_2O_3	15.48	22.36	Fe‴	47	66
FeO	11.53	10.58	Fe″	38	35
MnO	0.70	0.00	Mn	2	0
MgO	4.10	0.94	Mg	25	6
CaO	10.64	6.14	Ca	46	26
Na_2O	3.76	7.78	Na	29	59
K_2O	0.85	0.16	K	4	1
H_2O+	n.d.	n.d.			
H_2O-	2.12	1.16	O	600	600
Total.....	100.58	100.77			

	No. 8203	No. 1412K
Pleochroism		
X	pale bluish green	bluish green
Y	pale green	green
Z	yellowish green	yellowish green
	X > Y > Z	X > Y > Z
2V(+)	53°, 56°, 65°, 72°	74°, 75.5°
c Λ Z	60°	66°
α′	1.708 on (110)	1.721 on (110)
γ′	1.738 on (110)	n.d. on (110)

8203: Aegirinaugite crystal from syenite, No. 8203, middle reaches of Morotu River

1412K: Aegirinaugite core of aegirine crystal from syenite, No. 1412, headwaters of Morotu River

The chemical composition and optical properties of aegirinaugite No. 8203 from the syenite at the Morotu River are given in Table 5. Large crystals

TABLE 6.—AEGIRINE

Analyst: K. Yagi

	Wt. per cent				Atomic ratio		
	1411	1412S	1412H		1411	1412S	1412H
SiO_2	49.06	51.30	51.51	Si	191	199	19
TiO_2	2.34	0.72	0.68	Ti	7	2	
Al_2O_3	2.36	3.08	2.86	Al	11	14	1
Fe_2O_3	22.69	24.32	28.12	Fe‴	66	71	8
FeO	7.45	6.18	3.01	Fe″	24	20	1
MnO	0.68	0.08	0.00	Mn	2	0	
MgO	1.48	0.57	0.04	Mg	9	3	
CaO	4.80	3.08	1.80	Ca	20	13	
Na_2O	8.25	9.25	11.48	Na	62	69	8
K_2O	0.52	0.18	0.32	K	3	1	
H_2O+	n.d.	n.d.	n.d.				
H_2O-	0.71	0.78	0.24	O	600	600	60
Total...	100.34	99.54	100.06				

	No. 1411	No. 1412H
Pleochroism		
X	bluish green	pale straw ye low
Y	green	colorless
Z	greenish yellow	colorless
	X > Y > Z	X > Y = Z
2V(−)	66.5°	60°
c Λ Z	85°	89°
α′	1.748–1.771 on (110)	1.748–1.774 (110)
γ′	n.d.	n.d.

1411: Aegirine crystal from syenite No. 1411, headwaters of Morotu River

1412S: Aegirine crystal as a whole from syenite No. 1412, headwaters of Morotu River

1412H: Aegirine rim of aegirine crystal from syenite, No. 1412, headwaters of Morotu River

of aegirine in the pegmatitic syenite No. 1412 from the headwaters of the Morotu River show strong zoning with a core of aegirinaugite, a thick deep green shell of aegirine, and a pale outer rim of pure aegirine. Owing to the remarkable difference in color as well as the shape of cleaved flakes, it was possible to separate the different zones in the crushed powder under the binocular microscope, and a chemical analysis was made of the aegirinaugite core (1412K, Table 5), of the pure aegirine rim (1412H,

Table 6), and also of aegirine crystal as a whole (1412S, Table 6). The chemical composition and optic properties of the aegirinaugite No. 1412K are also given in Table 5. The aegirinaugite No. 1412K is much richer in Na_2O and Fe_2O_3, while poorer in CaO and MgO, than the aegirinaugite No. 8203 in which the rather high K_2O replaces a considerable portion of Na_2O. The chemical formulas of the aegirinaugites follow:

Aegirinaugite No. 1412K: $(Na, K, Ca)_{0.9}(Fe''', Fe'', Mg, Al)_{1.1}(Si, Ti, Al)_2O_6$

Aegirinaugite No. 8203: $(Na, K, Ca)_{0.8}(Fe''', Fe'', Mn, Mg, Al)_{1.2}(Si, Ti, Al)_2O_6$

Aegirine.—Aegirine is most abundant in syenites. In some pegmatitic facies crystals as large as 10 mm. are not uncommon. They are bounded by the forms (100), (010), (110), (101), and (111). In some cases aegirine is older than the feldspars, but usually aegirine forms anhedral crystals in the mesostasis between feldspars, suggesting that its formation was, for the most part, later than the formation of albite layers around the feldspars.

Besides these large crystals, aggregates of slender prismatic crystals or fine hairlike crystals radiating from a common center are found in the small drusy cavities. They are elongated along the c axis and attain 0.05 to 0.1 mm. in length. These small crystals belong to the last stage of crystallization and perhaps are the product of hydrothermal processes.

Large crystals are bluish green, grass green, or pale green. If they are zoned, the outer zones are always paler with increase of the aegirine molecule. The outermost zone is nearly pure aegirine and is very pale green or almost colorless. In the zoned crystals, therefore, the color deepens as it extends from the core until it attains maximum intensity, when it again diminishes to an almost colorless rim. Some thin needle crystals of aegirine are also nearly colorless. Parallel growth with arfvedsonite, aenigmatite, etc., is observed; small hematite flakes are sometimes included in the aegirine, showing the reaction relation between the two minerals found in the laboratory (Bowen, Schairer, Willems, 1930, p. 431). Though aegirine is generally fresh, some has been altered to a reddish-brown substance, which is probably iron hydroxide, and a yellow nontronitic mineral with high birefringence. Cleavage parallel to (110) is perfect. The chemical composition and optical properties of aegirines in the syenites at the Morotu River are given in Table 6. No. 1411 is an analysis of complete crystals showing slight zoning; No. 1412S is the bulk analysis of strongly zoned crystals; and No. 1412H is the pure aegirine rim of the same crystals. Optical properties also show wide variation.

Chemical formulas follow:

Aegirine No. 1411: $(Na, K, Ca)_{0.9}(Fe''', Fe'', Mn, Mg, Al)_{1.1}(Si, Ti, Al)_2O_6$

Aegirine No. 1412S: $(Na, K, Ca)_{0.8}(Fe''', Fe'', Mg, Al)_{1.1}(Si, Ti, Al)_2O_6$

Aegirine No. 1412H: $(Na, K, Ca)_{0.9}(Fe''', Fe'', Mg, Al)_{1.0}(Si, Ti)_2O_6$

Samples No. 1411 and No. 1412S are much alike; Sample No. 1412H is a pure aegirine richest in acmite molecule.

Amphiboles: Amphiboles in these rocks show wide variation. In the dolerites brown barkevikitic hornblende is intimately associated with titanaugite or titanbiotite. Kaersutite occurs abundantly in some monzonites and syenites. In these rocks kaersutite rarely is surrounded by a mantle of hastingsitic hornblende. Arfvedsonite is the only amphibole in most of the alkalic syenites, where it frequently occurs in parallel growth with aegirine. Aenigmatite, which is sometimes grouped in the amphibole family, occurs in some places in the monzonites and syenites. Common hornblende and rhombic amphiboles are absent. Only kaersutite and arfvedsonite were chemically analyzed; the other amphiboles were determined optically.

Barkevikite.—Brown hornblende is included in titanaugite or titanbiotite, or less often as isolated small crystals in some dolerites from Morotu Cape. In these rocks iron ore inclusions in titanaugite are often surrounded by a thin rim of hornblende, 0.01 to 0.1 mm. wide, which always has crystallographic orientation in common with the host augite crystals. Pleochroism is marked, but less intense than that of kaersutite.

X = pale violet brown Y = violet brown
Z = reddish brown $c\Delta Z = 2.5°$
$2V(-) = 80°$ (core) $\rightarrow 77°$ (margin)

By virtue of the above properties, this amphibole is clearly distinguished from common hornblende, but it is less titaniferous than kaersutite and is therefore designated barkevikite. As the two amphiboles do not occur in the same rock, the genetic relation of barkevikite and kaersutite is not certain.

Kaersutite.—Dark brown or almost black lustrous crystals of hornblende can be observed in the monzonite veins traversing the dolerite at Tyaki, along the Tiyo River. Chemical analysis of this hornblende shows unusually high titanium, by which the hornblende is determined as kaersutite, a rather rare variety of barkevikite. This mineral also occurs in the syenite from the triangulation point north to Tyaki.

Prismatic crystals, usually 0.5 to 3.0 mm. (and, rarely, 10 mm.) long, are bounded by the forms

(100), (010), with the basal pinacoid less prominent. Cleavage parallel to (110) is perfect, and twinning on (100) is common. Besides common inclusions, such as apatite, magnetite, or ilmenite, the crystals sometimes include featherlike light gray inclusions arranged diagonally in parallel rows. These inclusions have definite orientation relative to the cleavage of the host kaersutite crystals, but their extremely minute size makes identification impossible.

TABLE 7.—KAERSUTITE No. 1305G

Analyst: K. Yagi

	Wt. per cent		Atomic ratio
SiO_2	40.73	Si	604
TiO_2	8.47	Ti	94
Al_2O_3	10.62	Al	185
Fe_2O_3	1.39	Fe‴	16
FeO	11.91	Fe″	147
MnO	0.25	Mn	4
MgO	10.46	Mg	233
CaO	10.40	Ca	166
Na_2O	2.92	Na	84
K_2O	1.33	K	25
H_2O+	1.97	OH	194
H_2O-	0.23	O	2206
Total.....	100.68		

Pleochroism	
X	brownish yellow or straw yellow
Y	reddish brown
Z	deep chestnut brown with violet tinge
Absorption	Z > Y > X
Extinction angle	
c ∧ Z	16°, 17°, 19°
Optic angle	
2V(−)	77°, 78°, 79°, (in the core) → 73° (in the margin)
Index	
α'	1.680 on (110)
γ'	1.703 on (110)

Parallel growth with other mafic minerals is worthy of mention. The order of arrangement from the core to the margin is as follows:

Kaersutite → titanbiotite → sodic titanaugite
Kaersutite → sodic titanaugite → titanbiotite

In these cases, b and c axes are in common. The boundary between them is sharp and irregular. Magmatic resorption or an opacite margin, so common in the basaltic hornblende of the volcanic rocks, is absent.

The chemical composition and optical properties of the kaersutite from the monzonite No. 1305G at Tyaki are given in Table 7. During the chemical analysis the usual fusion by alkali carbonates failed to render all the material completely soluble in HCl, and therefore the fusion was repeated. Impurities in the silica remaining after the treatment by HF were unusually high and probably were largely TiO_2.

Kaersutite is a rare mineral, and there are so far only six known localities of analyzed kaersutite. They are Kaersut, Greenland—original locality (Lorenzen, 1886, p. 318; Gossner and Spielberger, 1930, p. 120; Drescher, 1933, p. 247); Linosa, Italy (Washington and Wright, 1908, p. 1870); Tikaisi, Oki Island, Japan (Tomita, 1934, p. 99); Uturyoto, Korea (Harumoto, 1933, p. 96); Yohodo, Korea (Kawano, 1934, p. 124); and Boulder Dam, U. S. A. (Campbell and Schenk, 1950, p. 684). When the chemical analyses of these are compared (Table 8), it is apparent that kaersutites can be classified into two groups—the one characterized by high Fe_2O_3 and low FeO, and the other by low Fe_2O_3 and high FeO, while the sum of iron oxides is rather constant throughout both groups. Kaersutite from Linosa, Yohodo, Uturyoto, and Boulder Dam belong to the first group, and they always occur in the basaltic rocks. Kaersutite from Kaersut and Morotu fall in the second group and occur in the leucocratic veins traversing the mafic rocks. Therefore, the difference in their chemical compositions might be ascribed to the difference in the physico-chemical environment of their parent rocks.

In the calculation of the chemical formula, the sum of Si + Al in the atom group is less than 8, when we take O + OH = 24, and therefore there should be replacement of Si-Ti in this kaersutite. Thus it is expressed as follows: $OH_{1.9}(Ca, Na, K)_{2.3}(Fe'', Fe''', Mn, Mg, Al)_{4.8}(Si, Ti, Al)_8O_{22}$. This is in agreement with Warren's general hornblende formula of $(OH, F)_2(Ca, Na, K, Mn)_{2-3}(Mg, Fe, Ti, Mn, Al)_5(Si, Al)_8O_{22}$ (Warren, 1930, p. 516).

Hastingsite.—Brownish-green hornblende is an essential mineral in some monzonite from the Morotu River. It is generally stout prismatic and occurs sometimes as isolated crystals, but usually in parallel growth with other mafic minerals. In some cases, soda-augite is surrounded by a thick mantle of this hornblende with the b and c axes in common, and the boundary is sharp. In other cases, kaersutite has narrow, incomplete titanbiotite fringes which are succeeded by hastingsite, the latter often coming into direct contact with the

kaersutite and forming parallel growth with the b and c axes in common. Zonal structure in hastingsite itself is sometimes observed. The color becomes more greenish, and both the extinction angle and the optic angle increase toward the margin. Optical properties follow:

Pleochroism:

X = pale brownish yellow—pale brownish green
Y = deep greenish brown—pale greenish brown
Z = deep greenish brown—bluish green

Absorption $Z > Y > X$

$2V(-) = 48°, 48.5°, 54°$ (core) $\rightarrow 64°$ (margin)
$c\Lambda Z = 5°$ (core) $\rightarrow 10° \rightarrow 20°$ (margin)

From these properties the hornblende should probably be regarded as hastingsite. The mode of zoning with other amphiboles shows that hastingsite is preceded by kaersutite, while it is earlier than arfvedsonite. Iwao (1939, p. 180) has mentioned that quasi-uniaxial hastingsite, which is always surrounded by arfvedsonite with gradual transition, occurs in some syenites and monzonites of this district.

TABLE 8.—KAERSUTITE

Analyst	Yagi	Teshima	Kawano	Usizima	Gossner	Washington	Gonyer
	(1)	(2)	(3)	(4)	(5)	(6)	(7)
SiO_2	40.73	39.78	38.30	39.20	39.50	40.85	41.46
TiO_2	8.47	7.00	6.02	6.53	10.33	8.47	5.70
Al_2O_3	10.62	14.13	12.87	13.87	11.12	9.89	14.24
Fe_2O_3	1.39	4.61	7.98	4.08	0.06	8.85	3.32
FeO	11.91	7.31	6.96	7.33	9.44	3.96	5.70
MnO	0.25	0.12	0.12	0.11	0.10	0.12	0.08
MgO	10.46	11.01	11.79	11.96	12.90	12.47	13.68
CaO	10.40	10.75	10.47	12.37	10.91	12.16	11.62
Na_2O	2.92	2.57	3.11	1.99	3.82	2.01	2.29
K_2O	1.33	1.58	1.30	1.45	1.43	0.63	1.72
H_2O+	1.97	0.37	1.10	0.87	0.59	0.19	0.12
H_2O-	0.23	0.22	0.26	0.27	—	—	—
F	—	—	0.05	—	—	0.28	0.42
	100.68	99.45	100.33	100.03	100.20	99.98*	100.35
Less 0 = F	—	—	0.02	—	—	0.12	0.18
Total.....	100.68	99.45	100.31	100.03	100.20	99.86	100.17

* Includes NiO, 0.10

1. From Tyaki, Morotu
2. From Tikaisi, Dogo, Oki, Japan (Tomita, 1934, p. 1)
3. From Yohodo, Korea (Kawano, 1934, p. 127)
4. From Uturyo Island, Korea (Harumoto, 1933, p. 100)
5. From Kaersut, Greenland (Gossner and Spielberger, 1930, p. 121)
6. From Monte Rosso, Linosa (Washington and Wright, 1908, p. 192)
7. From Boulder Dam, Arizona (Campbell and Schenk, 1950, p. 684)

Arfvedsonite.—Arfvedsonite is confined to alkali-rich monzonites or syenites with abundant aegirine. It occurs either in isolated crystals or in parallel growth with other mafic minerals. It is broad tabular or prismatic, deep bluish green with strong pleochroism. The size ranges between 0.02 mm. and 1 mm. but is usually 0.1 to 0.5 mm. Arfvedsonite sometimes occurs intergrown with aegirinaugite, aegirine, and hastingsite. Crystals of aegirine and arfvedsonite in parallel growth in the syenite No. 1412 from the headwaters of the Morotu River were separated from the parent rock. They were then crushed, and the arfvedsonite sample was carefully picked out under the binocular microscope. The chemical composition and optical properties of this arfvedsonite are given in Table 9.

Arfvedsonite in these alkalic rocks does not give perfect extinction on (010) plane between the crossed nicols, even in monochromatic light. Eskola and

Sahlstein (1930, p. 94) explained this phenomenon, occasionally observed in alkali hornblende, as due to the very fine perthitic exsolution lamellae. Sato (1936, p. 141) suggested that the optical abnormality

TABLE 9.—ARFVEDSONITE No. 1412
Analyst: K. Yagi

	Wt. per cent		Atomic ratio	
SiO_2	48.51	Si	756	
TiO_2	1.10	Ti	13	790
Al_2O_3	1.08	Al	21	
Fe_2O_3	17.53	Fe‴	204	
FeO	14.31	Fe″	186	429
MnO	0.66	Mn	8	
MgO	1.31	Mg	31	
CaO	5.43	Ca	91	
Na_2O	7.87	Na	237	358
K_2O	1.48	K	30	
H_2O+	1.48	OH	153	2400
H_2O-	0.17	O	2247	
Total.....	100.93			

Pleochroism	
X	greenish brown
Y	greenish blue
Z	deep greenish blue with violet tinge
Absorption	
Z > Y > X	
Extinction angle	
c Λ X	5°–6°
Index	
α'	1.686 on (110)
γ'	1.695 on (110)

of alkali amphibole in granite from Jehol, Manchuria, was presumably a result of its optical pseudo-activity. Iwao (1939, p. 183) remarked that the two pseudo-vibration directions of polarized light perpendicular to (010) do not intersect at right angles. This phenomenon was later discussed by Hori (1942) who concluded that the particles of the alkali hornblendes are arranged in random position in the (010) plane.

The analysis (Table 9) shows that CaO content is high, and total iron is also high, while MgO and Al_2O_3 are low. On the basis of OH + O = 24, the following formula is obtained: $OH_{1.5}(Na, K, Ca)_{3.6}(Fe'', Fe''', Mn, Mg)_{4.3}(Si, Ti, Al)_{7.9}O_{22.5}$. This is somewhat different from the accepted hornblende formula of $OH_2(Na, K, Ca)_{2-3}(Fe'', Fe''', Mg, Ti, Mn, Al)_5(Si, Al)_8O_{22}$. The group (Na, K, Ca) is too high, and the group (Fe″, Fe‴, Mg, Mn) is too low. However, if we accept the isomorphous replacement of Ca-Mg in this amphibole, just as in the pyroxene, we get the formula, $OH_{1.5}(Na, K)_{2.7}(Fe'', Fe''', Mn, Mg, Ca)_{8.2}(Si, Ti, Al)_{7.9}O_{22.5}$, which is in close agreement with the general hornblende formula.

Aenigmatite.—The crystallographic position of aenigmatite has not yet been settled, and it is sometimes omitted from the amphibole group (Fleischer, 1936, p. 346). In the present discussion, however, it is tentatively included.

Aenigmatite is a rather rare mineral and in this area is found only in some monzonites and syenites from the Morotu River. It seldom attains 1 mm. in size and is usually less than 0.1 mm. Cleavage parallel to (110) is well developed, and twinning is not observed. Pleochroism is extraordinarily strong and characteristic.

X = reddish brown
Y = deep chestnut brown
Z = dark brown or nearly opaque
Extinction angle $c\Delta Z = 27°–55°$
$\alpha' = 1.774$ on (110)

In the large crystals zonal structure is present; the core is deep greenish brown with the following strong pleochroism:

X = slightly greenish brown
Y = deep brown
Z = very dark brown to opaque

This pleochroism differs from that of aenigmatite and suggests that the mineral in the core is rhönite, though the exact optical properties could not be determined. Optical orientation is the same throughout the zoned crystals, but the boundary between the two minerals is very sharp. Therefore, it seems probable that rhönite and aenigmatite form a discontinuous reaction series, of which the earlier member is rhönite (Fleischer, 1936, p. 348). A similar relation was observed by Tomita (1934, p. 124) in the resorption products of kaersutite from Titaisi, Oki, Japan. It was not possible to obtain sufficient material for chemical analysis.

BIOTITE: Biotite is an important mafic mineral in all dolerites of the area, and in some monzonites also; but in syenites it is either subordinate in amount or absent. Biotite in the dolerites and monzonites is characterized by a deep reddish-brown color and strong pleochroism due to high content of titanium and should be called titanbiotite. The biotite in the syenites is a lepidomelanic variety characteristically deep brown with a greenish tinge.

Titanbiotite.—In dolerites titanbiotite forms small individual crystals or large crystals ophitically related to plagioclase. It is usually fresh, but may be altered to chlorite. The chemical composition and

TABLE 10.—TITANBIOTITE

Analyst: K. Yagi

	Wt. per cent			Atomic ratio					
	607	1305G		607		1305G			
SiO_2	33.60	36.12	Si	260		270			
TiO_2	5.79	8.27	Ti	33	425	46	424		
Al_2O_3	14.47	12.18	Al	132		108			
Fe_2O_3	7.00	2.21	Fe‴	41		13			
FeO	11.20	15.40	Fe″	73	250	96	245		
MnO	0.30	tr.	Mn	2		0			
MgO	11.56	12.15	Mg	134		136			
CaO	2.39	1.82	Ca	20		17			
Na_2O	0.41	1.15	Na	7	109	17	95		
K_2O	8.29	6.40	K	82		61			
H_2O+	3.12	4.03	OH	161	1200	200	1200		
H_2O-	2.00	0.96	O	1039		1000			
Total...	100.13	100.69							

	No. 607	No. 1305G
Pleochroism		
X	pale straw yellow	pale yellow or pale brownish yellow
Y = Z	dark reddish chestnut brown	dark reddish brown
	Y = Z > X	Y = Z > X
Optic angle		
2V(−)	0°, rarely 3°	0°, rarely 16°
Index		
γ	1.659	1.657

607: From dolerite, No. 607, Morotu Cape
1305G: From monzonite, No. 1305G, Tyaki, Tiyo Village

optical properties of titanbiotites in the dolerite (No. 607) at Morotu Cape and in the monzonite vein (No. 1305G) at Tyaki are given in Table 10. High titanium in both biotites is characteristic. The high CaO in the biotite No. 607 is partly due to impurity of titanaugite which could not be completely eliminated from the analyzed sample. However, the analyzed sample of biotite No. 1305G was very pure, and the CaO determined must be regarded as an ingredient of the biotite. Tsuboi, Sugi, and Iwao (1938, p. 453) have remarked that biotites of igneous origin have more or less CaO, while those in the metamorphic rocks have a very subordinate amount. The content of CaO in this biotite is the highest even among the biotites of igneous origin. Na_2O is also rather high, K_2O is accordingly low, and Na as well as Ca takes the place of K. Nockolds (1947, p. 401) discussed the relation between chemical composition and paragenesis in biotites of igneous origin. When the two biotites are plotted in a triangular diagram of MgO-FeO[2]-Al_2O_3, they fall within the field of biotites associated with hornblende, pyroxene, or olivine. This agrees with the actual paragenetic relation in these rocks. It is also noticed that Al_2O_3 content is low in these biotites, especially in the biotite No. 1305G, compared with biotites in the calcalkalic rocks.

The chemical formulas, computed on the basis of OH + O = 12, are:

Titanbiotite No. 1305G: $OH_2(K, Na, Ca)_{0.9}$ (Fe″, Fe‴, Mn, Mg, $Al)_{2.7}(Si, Ti, Al)_4O_{10}$

Titanbiotite No. 607: $(OH, O)_2(K, Na, Ca)_{1.1}$(Fe″,Fe‴, Mn, Mg, $Al)_{2.8}(Si, Ti, Al)_4O_{10}$

They are in close agreement with the general biotite formula, $(OH, F)_2(K, Na, Ca)_1$(Fe″,Fe‴,$Mg)_{2-3}$ Si_4O_{10}. In some biotite OH is less than 2, and the deficiency in OH is supplied by O. Prider (1939, p. 373) also has remarked that OH is usually lower than unity in the formula of titanium-rich biotites. In the present case, the titanbiotite No. 607 has a deficiency of OH, and the titanbiotite No. 1305G, which has higher titanium, has just enough OH to satisfy the formula.

Lepidomelanic biotite.—Titanbiotite is absent in nearly all the syenites, and its place is taken by small flakes of deep greenish-brown biotite, 0.05 to 0.3 mm. in size. Small radial aggregates of such biotite are found in some syenites, especially in close association with aegirine.

Pleochroism:

X = pale brownish yellow
Y = *Z* = dark chestnut brown with greenish tinge

Although no chemical analysis of this biotite was made, it is probably lepidomelanic biotite, rich in Fe_2O_3 and rather poor in TiO_2, judging from its characteristic color (Hall, 1941, p. 32). This inference is in accord with the bulk composition of the syenites.

IRON ORES: The abundance of iron ores and their remarkable association with other mafic minerals

[2] All Fe is expressed as FeO.

throughout the rock series are worthy of notice. Most of these ores are 0.1 to 0.5 mm. in size, but some attain 2 to 3 mm. They are square, prismatic, or sometimes skeletal. There is considerable variation in these ores, but their determination in the ordinary thin section is difficult. Study of polished sections[3] showed that there are at least four kinds of iron ores: magnetite, ilmenite hematite, and pyrite (listed in the order of their abundance). Iron ores in the dolerite No. 1507 from Takara Bridge show characteristic zonal arrangement. The inner part is composed of homogeneous hematite, while the thick shell consists of ilmenite, which has thin exsolution lamellae of an unknown mineral. Subordinate amounts of pyrite occur in small isolated grains. The cleanest sample of iron ore, which still contains some titanaugite, shows 26 per cent TiO_2, which shows that the bulk of the ore is ilmenite.

In the dolerites, titanbiotite often grows on iron ore, and barkevikite is sometimes formed as a reaction rim around the iron ores which are included in the titanaugite. When iron ore is included in olivine, however, no reaction rim is observed. In some syenites coronas of aegirine or arfvedsonite surround the iron ores included in soda-augite.

ANALCITE: Analcite is a rather constant constituent throughout the whole series. Beautiful crystals up to 10 mm. long are found in the miarolitic cavities of some monzonites at Takara Bridge. The crystals are snow white with pearly luster and are bordered chiefly by the form (211). Besides these euhedral crystals, analcitization of feldspar is common. It commences at the core and proceeds outward, eventually completely replacing the feldspar crystals. It is isotropic with $n = 1.490$, but very weak birefringence is sometimes observed with the aid of the gypsum plate.

TABLE 11.—ANALCITE NO. 901
Analyst: K. Yagi

	Wt. per cent		Atomic ratio	
SiO_2	55.37	Si	402	402
TiO_2	0.05	Ti	0	
Al_2O_3	22.73	Al	194	198
Fe_2O_3	0.73	Fe‴	4	
FeO	n.d.	Fe″	0	
MnO	0.08	Mn	0	
MgO	0.03	Mg	0	
CaO	0.82	Ca	7	175
Na_2O	10.93	Na	153	
K_2O	1.62	K	15	
H_2O+	8.38	H	406	406
H_2O-	0.10	O	1400	1400
Total.....	100.84			

Index of refraction n = 1.490

TABLE 12.—PREHNITE NO. 1509
Analyst: K. Yagi

	Wt. per cent		Atomic ratio	
SiO_2	43.84	Si	300	300
TiO_2	0.10	Ti	0	
Al_2O_3	22.56	Al	182	195
Fe_2O_3	2.52	Fe‴	13	
FeO	0.00	Fe″	0	
MnO	0.00	Mn	0	
MgO	0.25	Mg	2	199
CaO	26.39	Ca	194	
Na_2O	0.20	Na	2	
K_2O	0.10	K	1	
H_2O+	4.75	H	217	217
H_2O-	0.10	O	1200	1200
Total.....	100.81			

Optical properties
Colorless
Optic angle
2V(+) 70°
Index of refraction
α' 1.617 on (001)
γ' 1.635 on (001)

The chemical composition of clear analcite from the monzonite No. 901 at Takara Bridge is given in Table 11. The chemical formula is computed on the basis of O = 14, as follows: $H_{4.1}(Na, K, Ca)_{1.1}(Al, Fe''')_2Si_4O_{14}$. A notable amount of Na is replaced by K in this analcite, but not so much as in the case of potash analcite (Larsen and Buie, 1938, p. 838).

PREHNITE: Prehnite frequently fills the mesostasis between other minerals, or replaces feldspar in the dolerites or monzonites. It is usually 0.1 to 1.0 mm. in size, but kidney-shaped aggregates of radial fibrous crystals, 10 to 20 mm. in diameter, are found in the miarolitic druses in some monzonites. It is white or colorless and has perfect (001) cleavage. When feldspar is replaced by prehnite, the latter forms irregular radial fibrous aggregates in the center of feldspars. A pure sample of

[3] The writer is indebted to Mr. A. Sugaki of the Tohoku University for the determination of the iron ores on the polished sections.

prehnite, separated from the monzonite No. 1509 from Takara Bridge, was chemically analyzed (Table 12).

The high content of Fe_2O_3 is notable and presumably replaces Al_2O_3 isomorphously. It is interesting that prehnite is colorless in spite of this high Fe_2O_3. The chemical formula on the basis of O = 12, $H_2(Ca, Na, K, Mg)_2(Al, Fe'')_2Si_3O_{12}$, accords well with the formula of prehnite, $H_2Ca_2Al_2Si_3O_{12}$. Besides analcite and prehnite, some zeolites are present in subordinate amounts. Of these thomsonite and natrolite are common.

Accessory Minerals

APATITE: Apatite is a constant accessory in all rocks. It ranges from 0.05 to 0.5 mm. in length, and rarely attains 1 to 2 mm. It is colorless, free from minute pigmental inclusions, with $n_\omega = 1.635$. The mineral is probably fluorapatite.

SPHENE: Sphene, though present in all rocks of the area, is always smaller than 0.2 mm. across and occurs in small amounts only, even in rocks rich in titanium. It is usually a primary mineral, but it also occurs as an alteration product of ilmenite.

ZIRCON: Trifling amounts of zircon occur in some monzonites and syenites of the area. It always forms minute euhedral crystals.

PETROGRAPHY

General Description

The alkalic rocks of the Morotu district can be classified after Shand's (1947, p. 225) classification (Table 13). There are no over-saturated rocks in the district; all are either saturated or under-saturated and range from dolerites through monzonites to syenites, dolerites being by far the most abundant. No peraluminous rocks, characterized by the presence of biotite only, are found. Metaluminous rocks with both pyroxene and biotite are most common; some peralkalic rocks with predominant alkalic amphiboles and aegirine are also found, though the amount is small. There are no nepheline-bearing rocks, though some rocks carry as much as 8 per cent nepheline in the norm, and normative nepheline is represented mostly by the analcite in the mode. Whether or not such analcite rocks should be classified as feldspathoidal rocks is an interesting problem. Recently analcite has been considered a primary constituent. Some analcite rocks are regarded as a special case of nepheline rocks, the only difference being crystallization of analcite in place of nepheline (Lacroix, 1924, p. 529). In the Morotu rocks analcite is always interstitial between the feldspars and pyroxenes, and it is usually anhedral to other minerals, except in miarolitic cavities. Analcite also commonly replaces feldspar. These modes of occurrence suggest that the Morotu analcite is pneumatolytic rather early magmatic. Therefore, these analcite-bearing rocks are not here regarded as feldspathoidal rocks.

TABLE 13.—CLASSIFICATION OF THE MOROTU ALKALIC ROCKS

	p	m	s	k
Oversaturated	/	/	/	/
Saturated	/	Monzonite	Syenite	Syenite
Undersaturated				
Feldspathoidal	/	/	/	/
Nonfeldspathoidal	/	Dolerite	/	/

Dolerites

Dolerites are the most abundant rock type. They are the only type present in smaller sheets or dikes and constitute the main part of the larger intrusive bodies.

All the dolerites are a metaluminous type, characterized by the mineral association of titanaugite, titanbiotite, and hastingsite or barkevikite. They vary widely in their appearance from fine-grained basaltic rocks to coarse-grained gabbroic rocks, but there is no glassy facies. Fine-grained rocks are confined to the chilled contact with sediments. Porphyritic texture is rarely observed in them. Most of the dolerites show well-developed ophitic texture under the microscope; some coarse-grained ones are granular.

In porphyritic basaltic rocks, the phenocrysts are plagioclase (An_{55-65}), diopsidic augite, and olivine (sometimes microphenocrysts). The groundmass consists of plagioclase (An_{50}), anorthoclase, titanaugite, titanbiotite, iron ores, and accessory minerals. Chlorite, calcite, and other secondary minerals occur, replacing the

TABLE 14.—CHEMICAL ANALYSES AND NORMS OF DOLERITES

Analyst: K. Yagi

	(1)	(2)	(3)	(4)	(5)	(6)
SiO_2	46.38	46.54	46.62	46.64	48.23	50.18
TiO_2	2.24	2.61	2.59	1.56	2.05	1.07
Al_2O_3	15.86	16.08	16.28	17.50	16.41	17.73
Fe_2O_3	4.02	3.55	6.29	2.34	3.88	1.54
FeO	6.29	7.95	5.40	7.32	7.17	3.43
MnO	0.08	0.18	0.04	0.16	0.20	0.14
MgO	7.30	6.97	4.51	6.66	6.00	6.15
CaO	8.68	9.81	9.12	7.86	7.24	14.70
Na_2O	2.99	2.03	4.25	3.14	4.42	3.12
K_2O	1.64	1.42	1.79	1.30	0.90	0.70
H_2O+	3.00	2.16	3.62	3.80	3.24	1.45
H_2O-	0.36	0.59	0.10	0.68	0.22	0.45
P_2O_5	0.24	0.25	0.30	0.20	0.16	0.19
CO_2	1.01	n.d.	n.d.	n.d.	n.d.	n.d.
Total..	100.39	100.14	100.91	99.16	100.12	100.85
Or	9.45	8.34	10.56	7.78	5.56	3.89
Ab	25.15	17.29	25.15	26.72	35.63	30.44
An	25.30	30.58	20.02	29.47	22.24	32.53
Ne	—	—	5.96	—	0.85	3.12
Di Wo	6.73	6.84	9.86	3.60	5.34	16.59
Di En	4.90	4.40	8.10	2.10	3.40	12.20
Di Fs	1.19	1.98	0.53	1.32	1.58	2.77
Hy En	0.80	9.60	—	1.60	—	—
Hy Fs	0.13	4.09	—	0.79	—	—
Ol Fo	8.82	2.38	2.24	9.10	8.12	2.24
Ol Fa	2.65	1.22	0.10	5.30	4.28	0.61
Mt	5.80	5.34	9.05	3.48	5.57	2.32
Il	4.26	5.02	4.86	3.04	3.95	1.98
Ap	0.67	0.67	0.67	0.34	0.34	0.34
Or	15.7	14.8	18.9	12.2	8.8	6.8
Ab	42.0	30.8	45.2	41.8	56.1	36.0
An	42.3	54.4	35.9	46.0	35.1	57.2
Wo	48.9	25.3	53.3	38.2	51.7	52.6
En	41.5	52.2	43.8	39.3	32.9	38.6
Fs	9.6	22.5	2.9	22.5	15.4	8.8

1. No. 605, Morotu Cape
2. No. 82, Headwaters of the Morotu River
3. No. 1507, Takara Bridge, Tiyo Mine
4. No. 607, Morotu Cape
5. No. 55D, Headwaters of the Morotu River
6. Aggregate No. 1305K, Tyaki, Tiyo Village

original minerals, and fresh samples are difficult to obtain.

The minerals composing the dolerites of ophitic texture are plagioclase, anorthoclase, olivine, titanaugite, barkevikite, titanbiotite, iron ore, and accessory minerals. Analcite and other secondary minerals are also very widely distributed. Plagioclase (An_{40-60}) is usually intergrown with titanaugite, showing remarkable ophitic texture. It is occasionally clouded by secondary alteration products. Anorthoclase occurs only as a thin rim around plagioclase, or as mesostasis, and is always quite clear. Olivine is hyalosiderite with rather high fayalite content (Fa_{37-50}) and is nearly always replaced by chlorite or some other secondary minerals. Titanaugite and titanbiotite are the most abundant mafic minerals and are rather fresh in nearly all rocks. Barkevikite is usually subordinate. It is most abundant in the soda-rich dolerite No. 55D which is closely associated with soda-rich syenite in the Morotu River. This rock is characterized by sodic plagioclase (An_{35-40}) and barkevikite without titanbiotite.

Some dolerites have numerous pea-shaped aggregates 1 to 4 cm. in diameter, which are rather leucocratic compared to the dark host dolerites. These aggregates project from the weathered surface because of their greater resistance. Such aggregates consist of plagioclase (An_{54-62}), olivine, diopsidic augite, with a small amount of anorthoclase, barkevikite, sphene, apatite, and iron ores.

The chemical compositions of dolerites from several localities are given in Table 14. The analyzed specimens are described briefly:

DOLERITE NO. 605: Morotu Cape. Texture: medium-grained, ophitic. Main components: plagioclase An_{60}, anorthoclase, titanaugite, titanbiotite, olivine, iron ore. Minor components: barkevikite, sphene, analcite, apatite.

DOLERITE NO. 607: Morotu Cape. Texture: coarse-grained, ophitic. Main components: plagioclase An_{54}, anorthoclase, titanaugite (*see* Table 3), titanbiotite (*see* Table 10), olivine, iron ore. Minor components: barkevikite, analcite, sphene, apatite.

DOLERITE NO. 82: Headwaters of the Morotu River. Texture: fine-grained, ophitic. Main components: plagioclase An_{64}, anorthoclase, titanaugite, titanbiotite, olivine (Fa_{37-50}). Minor components: barkevikite, analcite, apatite.

DOLERITE NO. 1507: Takara Bridge, Tiyo. Texture: coarse-grained, ophitic. Main components: plagioclase An_{40}, anorthoclase, titanaugite (*see* Table 3), titanbiotite, olivine, iron ore, analcite. Minor components: barkevikite, sphene, apatite.

DOLERITE NO. 55D: Middle reaches of the Morotu River. Texture: slightly porphyritic with micro-

phenocrysts in ophitic matrix. Main components: plagioclase An_{35-40}, anorthoclase, barkevikite, titanaugite, olivine. Minor components: iron ore, analcite, sphene, apatite.

AGGREGATE IN DOLERITE NO. 1305K: Tyaki, Tiyo Village. Texture: ophitic. Main components: plagioclase An_{54-62}, diopsidic augite, olivine. Minor components: anorthoclase, barkevikite, iron ore, titanbiotite, sphene, apatite, analcite.

The chemical compositions of these rocks are very similar in spite of their different appearance. Alkalis, especially Na_2O and iron oxides, are rather high, and MgO and CaO are rather low. TiO_2 is high in all rocks. The dolerite No. 55D is characterized by higher Na_2O and lower K_2O compared with other rocks, and this peculiarity is well represented in the acidic composition of plagioclase, abundance of barkevikite, and absence of biotite. The aggregate from dolerite No. 1305K shows very high CaO and very low iron and TiO_2. Alkalis are also rather low. This is in agreement with the presence of diopsidic augite instead of titanaugite. An igneous rock with the chemical composition of this aggregate has not been found elsewhere.

Monzonites

Monzonites always occur in the inner parts of the larger intrusive bodies or as irregular veins or schlieren within the dolerites. They are usually more leucocratic and coarser-grained than dolerites and sometimes even pegmatitic. Abundant miarolitic cavities are present in almost all monzonites. The texture of the monzonites is always hypidiomorphic granular, never ophitic as in dolerites. Though mineral compositions are variable, all monzonites in the district fall in the metaluminous type. The feldspar is remarkably zoned from plagioclase core to anorthoclase or microperthite shell with a thin albite rim. The plagioclase is oligoclase (An_{20-30}) and is generally fairly fresh. In most cases the shell is microperthite with rather thin lamellae, but in some rocks it is homogeneous anorthoclase. The pyroxene covers a wide range in composition from soda-augite to aegirine, and sometimes there is also titanaugite. Titanbiotite is much less abundant than in the dolerites. The amphiboles, which occur in small amounts, are kaersutite, hastingsite, or arfvedsonite. Barkevikite and aenigmatite are seldom present. Accessory minerals and secondary minerals are similar to those in the dolerites. Chemical compositions of some monzonites are shown in Table 15, and petrographic description of these rocks follows:

TABLE 15.—CHEMICAL ANALYSES AND NORMS OF MONZONITES

Analyst: K. Yagi

	(1)	(2)	(3)	(4)
SiO_2	53.20	56.10	56.14	58.21
TiO_2	1.82	1.39	1.43	2.08
Al_2O_3	17.81	18.77	17.03	18.16
Fe_2O_3	3.08	3.19	5.14	1.99
FeO	4.96	2.57	4.33	3.30
MnO	0.10	0.17	0.07	0.07
MgO	1.62	1.22	1.38	1.76
CaO	4.52	3.98	4.02	3.41
Na_2O	6.56	6.49	5.77	6.99
K_2O	2.19	3.39	2.97	2.77
H_2O+	2.84	1.76	1.39	1.11
H_2O-	0.47	0.83	0.14	0.48
P_2O_5	0.43	0.41	0.34	0.11
Total.....	99.60	100.27	100.15	100.44
Q	—	—	1.98	—
Or	12.79	20.02	17.79	16.68
Ab	48.73	49.78	48.73	56.59
An	12.79	11.95	11.68	9.73
Ne	3.69	2.84	—	1.42
Di Wo	2.90	2.20	2.67	2.67
Di En	1.50	1.80	1.80	1.90
Di Fs	1.32	0.13	0.66	0.53
Ol Fo	1.82	0.84	1.70	1.75
Ol Fa	1.84	—	0.79	0.41
Mt	4.41	4.64	7.42	3.02
Il	3.50	2.58	2.74	3.95
Ap	1.01	1.01	0.67	0.34
Or	17.3	24.5	22.8	20.1
Ab	65.6	60.9	62.3	68.2
An	17.1	14.6	14.9	11.7
Wo	50.7	53.3	35.0	52.3
En	26.2	43.6	45.9	37.3
Fs	23.1	3.1	19.1	10.4

1. No. 902, Takara Bridge, Tiyo Mine
2. No. 901, Takara Bridge, Tiyo Mine
3. No. 1413, Headwaters of the Morotu River
4. No. 1305G, Tyaki, Tiyo Village

MONZONITE NO. 901: Takara Bridge, Tiyo Mine. Texture: coarse-grained, hypidiomorphic granular. Main components: plagioclase An_{22}, microperthite,

soda-augite, aegirinaugite, analcite (*see* Table 11). Minor components: aenigmatite, arfvedsonite, titanbiotite, iron ore, chlorite, calcite.

MONZONITE NO. 902: Takara Bridge, Tiyo Mine. Texture: coarse-grained, hypidiomorphic granular. Main components: plagioclase An_{30}, microperthite, soda-augite (*see* Table 4), aegirinaugite, titanbiotite, analcite. Minor components: iron ore, chlorite, arfvesdonite.

MONZONITE NO. 1413: Headwaters of the Morotu River. Texture: fine-grained, hypidiomorphic granular. Mafic minerals form aggregates. Main components: plagioclase, microperthite, soda-augite, aegirinaugite, hastingsite. Minor components: titanbiotite, iron ore, arfvedsonite, sphene, zircon, analcite.

MONZONITE NO. 1305G: Tyaki, Tiyo Village. Texture: very coarse-grained, with abundant miarolitic cavities, hypidiomorphic granular. Main components: microperthite, plagioclase An_{28}, kaersutite (*see* Table 7), titanbiotite (*see* Table 10), titaniferous soda-augite (*see* Table 4), aegirinaugite. Minor components: iron ore, analcite, apatite, sphene.

The alkali content of the monzonites is much higher and MgO and CaO much lower than in the dolerites. The amount of Fe_2O_3 is almost the same as that of FeO. The high TiO_2 content in the monzonite No. 1305G is in agreement with the abundance of kaersutite and titanbiotite, both of which are present only in small amount in other monzonites.

Syenites

Syenites are confined to the inner parts of the intrusive bodies and are usually accompanied by monzonites. They are all very leucocratic, and the grain size varies from fine to coarse, sometimes even pegmatitic. They are also very rich in minute miarolitic cavities. A lower color index and absence of zoned feldspar, characteristically present in the monzonites, distinguish the syenites from monzonites, though the distinction in the field is sometimes difficult.

Mineralogically syenites are classified into peralkalic syenites—characterized by arfvedsonite, hastingsite, and aenigmatite—and subaluminous syenites, which lack these alkali amphiboles. The feldspar is mostly microperthite, which is much coarser than in the monzonites, and oligoclase, which is very subordinate or rare. Homogeneous anorthoclase is not found. Aegirine and aegirinaugite are the most important mafic minerals. Barkevikite and kaersutite are very rare. Titanbiotite is usually absent, and its place is taken by some lepidomelanic biotite, though the latter is present only in small amount. Some of the representative specimens were chemically analyzed (Table 16).

The syenite at Takara Bridge (No. 1501) has 8 per cent normative nepheline, but no modal nepheline. Therefore, this high normative nepheline is evidently a result of the abundance of analcite in miarolitic cavities or replacing feldspars. The syenite No. 1310 at the top of the triangulation point north of Tyaki differs from the rest of the syenites in its abundance of kaersutite and titanbiotite. Unlike most of the syenites, syenite No. 55S, forming a thin vein in the dolerite in the middle reaches of the Morotu River, consists of microperthite, albite, soda-augite, and rare lepidomelanic biotite. Chemically the rock is characterized by the great predominance of Na_2O over K_2O.

SYENITE NO. 1410: Headwaters of the Morotu River. Texture: coarse-grained, hypidiomorphic granular. Main components: plagioclase An_{23}, microperthite, soda-augite, aegirinaugite, aegirine, analcite. Minor components: lepidomelanic biotite, iron ore, apatite, sphene.

SYENITE NO. 8302: Middle reaches of the Morotu River. Texture: coarse-grained, hypidiomorphic granular. Main components: microperthite (*see* Table 2), plagioclase An_{30}, aegirinaugite (*see* Table 5), soda-augite, aegirine. Minor components: analcite, arfvedsonite, hastingsite, lepidomelanic biotite, iron ore.

SYENITE NO. 1411: Headwaters of the Morotu River. Texture: coarse hypidiomorphic granular. Main components: microperthite, plagioclase An_{16}, aegirine (*see* Table 6), aegirinaugite, arfvedsonite. Minor components: hastingsite, lepidomelanic biotite, analcite, iron ore.

SYENITE NO. 1412: Same locality as No. 1411. Texture: coarse hypidiomorphic granular. Main components: microperthite (*see* Table 2), plagioclase An_{16}, aegirine (*see* Table 6), aegirinaugite (*see* Table 5), arfvedsonite (*see* Table 9). Minor components: analcite, aenigmatite, lepidomelanic biotite, hastingsite, iron ore, apatite.

SYENITE NO. 1310: Triangulation point 481.4 m. north of Tyaki, Tiyo Village. Texture: coarse hypidiomorphic granular. Main components: microperthite, oligoclase, aegirine, aegirinaugite, titan-

SiO_2	
TiO_2	
Al_2O_3	
Fe_2O_3	
FeO	
MnO	
MgO	
CaO	
Na_2O	
K_2O	
H_2O+	
H_2O-	
P_2O_5	
Total....	
Q	
Or	
Ab	
An	
Ne	
C	
Ac	
Di	Wo
	En
	Fs
Hy	En
	Fs
Wo	
Ol	Fo
	Fa
Mt	
Il	
Hm	
Ap	
Or	
Ab	
An	
Wo	
En	
Fs	

1. No. 1
2. No. 1
3. No. 8
4. No. 1
5. No. 5
6. No. 1
7. No. 1

TABLE 16.—CHEMICAL ANALYSES AND NORMS OF SYENITES

Analyst: K. Yagi

	(1)	(2)	(3)	(4)	(5)	(6)	(7)
SiO_2	56.86	59.39	59.88	60.39	60.40	61.03	61.31
TiO_2	0.53	0.80	1.21	0.98	0.81	0.67	0.89
Al_2O_3	20.19	15.39	18.98	17.85	18.34	15.52	19.44
Fe_2O_3	3.63	6.43	3.51	3.84	3.72	5.08	1.75
FeO	1.40	3.28	2.58	2.36	2.53	2.98	1.89
MnO	0.13	0.18	0.07	0.03	0.15	0.07	0.06
MgO	0.33	0.56	0.37	0.31	1.12	0.47	0.39
CaO	1.64	2.18	2.27	2.79	1.96	1.87	1.17
Na_2O	7.42	7.08	6.11	6.65	8.32	6.19	7.03
K_2O	4.39	3.78	4.09	3.14	0.84	5.15	4.95
H_2O+	3.29	1.00	1.40	1.06	1.43	0.83	1.63
H_2O-	0.07	0.17	0.08	0.24	0.41	0.39	0.20
P_2O_5	0.07	0.10	0.25	0.52	0.50	0.13	0.13
Total.....	99.95	100.34	100.80	100.16	100.53	100.38	100.84
Q	—	0.90	4.14	5.16	3.90	2.70	—
Or	26.13	22.24	24.46	18.35	5.00	3.058	29.47
Ab	47.67	58.16	51.35	56.07	70.21	50.83	56.59
An	8.06	—	9.45	9.73	7.23	—	5.00
Ne	7.95	—	—	—	—	—	1.70
C	0.31	—	1.02	—	1.12	—	0.61
Ac	—	1.85	—	—	—	1.39	—
Di Wo	—	1.97	—	0.35	—	2.09	—
Di En	—	1.40	—	0.30	—	1.20	—
Di Fs	—	0.40	—	—	—	0.79	—
Hy En	—	—	0.90	0.50	2.80	—	—
Hy Fs	—	—	—	—	0.53	—	—
Wo	—	2.20	—	—	—	1.39	—
Ol Fo	0.56	—	—	—	—	—	0.70
Ol Fa	—	—	—	—	—	—	0.51
Mt	3.25	8.35	5.10	4.87	5.34	6.73	2.55
Il	1.06	1.52	2.28	1.82	1.52	1.22	1.67
Hm	1.44	—	—	0.48	—	—	—
Ap	—	0.34	0.67	1.34	1.01	0.34	0.34
Or	31.9	27.7	28.7	21.8	6.1	37.6	32.4
Ab	58.3	72.3	60.2	66.6	85.1	62.4	62.1
An	9.8	0	11.1	11.6	8.8	0	5.5
Wo	—	52.3	0	30.4	0	63.6	—
En	—	37.1	100	69.6	84.1	21.9	—
Fs	—	10.6	0	0	15.9	14.5	—

1. No. 1501, Takara Bridge, Tiyo Mine
2. No. 1411, Headwaters of the Morotu River
3. No. 8203, Middle reaches of the Morotu River
4. No. 1410, Headwaters of the Morotu River
5. No. 55S, Headwaters of the Morotu River
6. No. 1412, Headwaters of the Morotu River
7. No. 1310, Triangulation point 481.4 m. north of Tyaki, Tiyo Village

biotite, kaersutite. Minor components: aenigmatite, iron ore, analcite, apatite, sphene.

SYENITE NO. 55S: Middle reaches of the Morotu River. Texture: fine hypodiomorphic granular, Main components: plagioclase An_{0-5}, microperthite-soda-augite. Minor components: lepidomelanic biotite, iron ore, calcite.

SYENITE NO. 1501: Takara Bridge, Tiyo Mine. Texture: very fine-grained hypidiomorphic granular. Main components: microperthite, anorthoclase (2V = 60°,68°,70°,72°), albite, aegirine, aegirinaugite, analcite. Minor components: lepidomelanic biotite, iron ore.

PETROCHEMISTRY

Bulk Composition

Chemical composition of all rock types (Tables 14, 15, 16) shows a wide gap in SiO_2 content; dolerites range from 46 to 48 per cent, and monzonites and syenites from 53 to 61 per cent. The lack of a representative between these two groups poses an interesting problem. A gap like this has been found in certain other rock series, both alkalic and calcalkalic.

A normal variation diagram of the Morotu rocks, constructed by plotting the weight percentage of the oxides against that of SiO_2, shows marked scattering of projection points. High fractionation during the crystallization, which is apparent from complex mineral assemblages, is probably responsible for this marked scattering. Contents of Al_2O_3 and Fe_2O_3, although they are rather constant, show a reciprocal relation; high Fe_2O_3 is accompanied by low Al_2O_3 in some of the peralkalic syenites. FeO decreases toward the felsic end, but less rapidly than MgO. Both alkalis are decidedly high, but the Na_2O curve is convex upward, while that for K_2O is concave, attaining its highest value, 5 per cent, at the felsic end. The alkali-lime index is 50.0, placing the assemblage in the "alkalic series" as defined by Peacock (1931, p. 57). High TiO_2 throughout the series is notable. H_2O+ decreases from about 3 per cent in the dolerites to about 1 per cent in the syenites.

A triangular diagram (devised by Larsen, 1938, p. 516) of the Morotu rocks is shown in Figure 8. The average composition of Japanese volcanic rocks (Yamada, 1930, p. 1–6) is also given for comparison. The lines for Morotu rocks are nearly horizontal, whereas those of the Japanese volcanic rocks, which are typically calcalkalic, dip steeply downward to the right. This difference emphasizes the alkalic character of the present series.

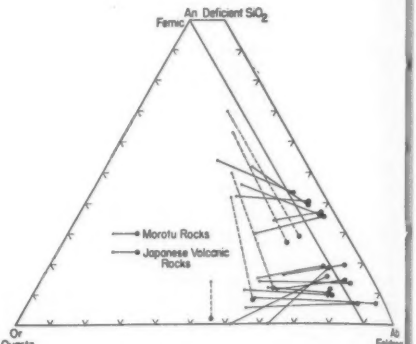

FIGURE 8. LARSEN TRIANGULAR DIAGRAM OF THE MOROTU ROCKS
With Japanese volcanic rocks for comparison. Weight per cent

In the normative compositions (Tables 14, 15, 16), normative quartz, which is limited to the syenites, is either absent or scarce. Normative nepheline appears in nearly half the analyses and sometimes amounts to 8 per cent in spite of the absence of nepheline in the mode. Its presence in the norm may for the most part be attributed to the abundance of analcite, but other factors which will be given later are also responsible. Normative acmite is rather restricted in spite of the occurrence of aegirine-augite, aegirine, or arfvedsonite. Excess of normative over modal olivine in the dolerites, as well as the presence of normative olivine in the olivine-free monzonite (No. 1305G) or syenite (No. 1310), can be explained by the abundant presence of titanbiotite, which yields much olivine in the normative calculation. Abundant normative ilmenite is also characteristic of this rock series.

Crystallization of Rock-Forming Minerals

FELDSPAR: Normative feldspar compositions of the Morotu rocks are projected in the triangular diagram Or-Ab-An (Fig. 9). Although the projected points appear to be scattered, a

trend is apparent upon close inspection. The feldspar in the dolerites (near labradorite) become increasingly enriched in the Ab molecule, while the Or molecule remains rather constant, and in the later stage (syenites) there is a great enrichment in the Or molecule at the expense of the Ab molecule, ending at about $Or_{37}Ab_{63}$.

A second, minor trend is suggested by the points representing soda-rich dolerite No. 55D and syenite No. 55S. This trend is dominantly controlled by increase of the Ab molecule, the percentage of Or molecule remaining nearly constant, and finds support in three of the analyzed samples of plagioclase plotted in Figure 10.

Bulk composition of the four analyzed feldspars and the estimated composition of each alkali feldspar (microperthite) and plagioclase are shown in Figure 10. Circles representing the bulk composition are joined by lines to the points representing the plagioclase and the microperthite in the same rocks. The two trends observed in the normative feldspars are also given in this diagram. Compositions of bulk feldspars Nos. 902, 8203, and 1412 all lie on the lower part of the main course of normative feldspars, and it thus seems likely that the trend of normative feldspars shows the course of evolution of the bulk feldspar composition in this rock series. Feldspar No. 1219, however, has a composition entirely different from that of the other feldspars in that its Or molecule predominates. At present the writer cannot find any explanation for this. In none of these cases did the feldspars crystallize out of a homogeneous single phase; they always consisted of plagioclase (P_1, P_2, P_3) and microperthite (A_1, A_2, A_3), although in the syenites the amount of plagioclase is very small (*cf.* Tuttle, 1952, p. 107).

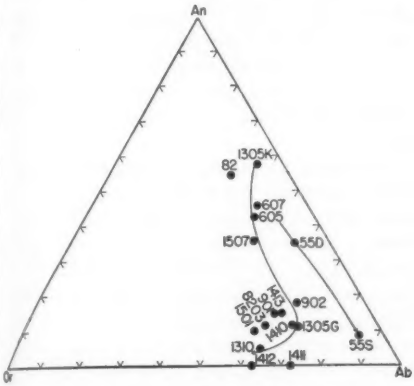

FIGURE 9. NORMATIVE FELDSPAR DIAGRAM OF THE MOROTU ROCKS
Weight per cent

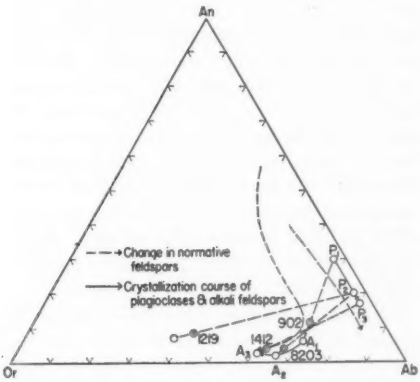

FIGURE 10. CRYSTALLIZATION COURSE OF THE PLAGIOCLASE AND ALKALIC FELDSPARS IN THE MOROTU ROCKS
Weight per cent

The trend of the crystallization of feldspars is given in Figure 10. Plagioclase (about An_{55}) crystallized out first from the parental magma and formed a continuous reaction series up to about An_{15}. Following the crystallization of plagioclase, alkali feldspar separated out as a single phase of anorthoclase in the dolerites, forming mantles around plagioclase. Anorthoclase seems quite homogeneous; no exsolution perthite was observed. As these anorthoclases have not been chemically analyzed, they are not plotted in the diagram. In the later stage, when monzonites were formed, plagioclase assumed the composition P_1, and alkali feldspar, of composition A_1, crystallized out as a separate phase of microperthite, giving the mixture of feldspars, whose bulk composition is shown by point 902 in the diagram.

In the more advanced stage, the plagioclase became more albitic, and the alkali feldspar was enriched in the Or molecule. In syenites

richest in alkalis, such as No. 1412, plagioclase is An_{16}, and microperthite is $Or_{61}Ab_{37}An_2$.

In summary the main crystallization course of the feldspars in the Morotu rocks is stated as follows: plagioclase crystallized out first, gradually became richer in the Ab molecule, and was succeeded by the alkali feldspar. The latter was anorthoclase, rich in Ab molecule, and became richer in Or molecule, trending toward the minimum point in the Or-Ab system of about $Ab_{65}Or_{35}$. In the dolerites formed at higher temperatures, the anorthoclase remained homogeneous, but, in the monzonites and syenites where the temperatures were much lower, feldspars are represented by microperthite. The width of these microperthite lamellae, which are very coarse in some pegmatitic syenites, seems to be dependent on the temperatures.

This behavior agrees with the experimental findings of Bowen and Tuttle (1950, p. 499), who showed that alkali feldspars homogeneous at higher temperatures unmix with lowering temperatures into potash feldspar and soda feldspar—probably very pure orthoclase and albite. Bowen and Tuttle determined the maximum point of the solvus curve at 660° ± 10°, and at a composition close to Ab_{55}.

Alkali feldspars in the trachytes and associated lavas of the Saisyu Island, Korea (Haraguchi, 1930, p. 539; Tomita, 1932, p. 507), also show an enrichment in Or molecule in the later stage. In many alkalic rocks of other regions, however, the reverse trend prevails. Barth (1937, p. 186–189) mentioned that alkali feldspars in volcanic as well as pegmatitic rocks form a reaction series from potassic to sodic, comparable to the reaction series of plagioclase feldspars. Feldspar in the felsic rocks of Oki, Japan, proceeded from potash feldspar to soda anorthoclase (Tomita, 1932, p. 151). A similar trend of feldspars has recently been observed by Oftedahl (1948, p. 52) in the rocks of the Oslo region.

The apparent discrepancy between the two feldspar trends, one of Or enrichment, and the other of Ab enrichment, can be explained by the difference of the position of the bulk feldspar composition relative to a trough which extends from the minimum point of the Or-Ab join toward the inner part of the Or-Ab-An system. In many cases, feldspar composition of the rocks lies to the Or side of this trough, and therefore, as the alkali feldspars proceed to this minimum, they are enriched in Ab molecule. On the other hand, the feldspar composition of the Morotu and Saisyu rocks lies on the Ab side of the trough, and crystallization toward the minimum results in enrichment of Or molecule in the alkali feldspars. The rarity of this trend toward potash enrichment should be ascribed to the fact that few rocks have a bulk feldspar composition like that of the Morotu rocks.

Sometimes feldspars in this area show a continuous reaction series between plagioclase and alkali feldspar. Mountain (1925, p. 338) established a continuous solid-solution series from andesine ($Ab_{65}An_{35}$) to anorthoclase ($Ab_{65}Or_{35}$). Occurrence of such potash andesine or potash oligoclase has been described from many basaltic rocks of Japan and China (Tomita, 1932, p. 153; 1933, p. 7) and from the Hawaiian Islands (Macdonald, 1942, p. 800). Recently Kuno (1950, p. 967) also found a continuous variation from andesine to potash oligoclase and finally to anorthoclase in the andesitic rocks and their xenoliths in the Hakone volcano, Japan. In the Morotu rocks, however, the feldspars occur always in two phases of plagioclase and alkali feldspar, instead of one single phase, and there is no solid-solution series between these two feldspars.

PYROXENE: The following pyroxene molecules are generally accepted and are used in the present discussion:

Wollastonite	$CaSiO_3$	Wo
Ferrosilite	$FeSiO_3$	Fs
Jadeite	$NaAlSi_2O_6$	Jd
Enstatite	$MgSiO_3$	En
Acmite	$NaFeSi_2O_6$	Ac
Sodium metasilicate	Na_2SiO_3	Ns

Tschermak's molecule $(Mg, Fe)Al_2SiO_6$ and the molecule $Fe_2O_3 \cdot 3SiO_2$ or $Al_2O_3 \cdot 3SiO_2$, suggested by Washington and Merwin (1927, p. 236), are omitted because of their hypothetical nature and because of the recent experiments on artificial melts. There are several ways to calculate the molecular compositions of pyroxenes (Washington and Merwin, 1927, p. 237; Tomita, 1937, p. 354). The writer suggests a satisfactory and simple method, similar to the

calculation of the norm, as follows (all calculation in molecular proportion):

1. (a) K_2O is added to Na_2O, and MnO to FeO. An amount of Fe_2O_3 equal to Na_2O is allotted for acmite.

(b) If $Na_2O > Fe_2O_3$, an excess of Na_2O is allotted to Al_2O_3 for jadeite.

(c) If $Na_2O > Fe_2O_3 + Al_2O_3$, an excess of Na_2O is reserved for Ns.

(d) If $Na_2O < Fe_2O_3 + Al_2O_3$, an excess of Al_2O_3 or an excess of Fe_2O_3 and of Al_2O_3 is kept as remainder.

2. (a) CaO is allotted to FeO and MgO for Hd and Di respectively; the relative proportion of the two determined as FeO and MgO occurs in the analysis.

(b) If CaO > FeO + MgO, an excess CaO is reserved for Wo.

(c) If CaO > FeO + MgO, excess FeO and MgO are reserved for Fs and En, respectively.

3. (a) Allot the necessary amount of SiO_2 to the above-mentioned oxides to complete the pyroxene molecules.

(b) Excess of SiO_2, Al_2O_3, Fe_2O_3, and TiO_2 in the above calculation is kept aside as remainder composition.

(c) If there is a deficiency of SiO_2, TiO_2 is added to SiO_2. If there is still deficiency, Al_2O_3 is added to SiO_2. When there is further deficiency of SiO_2, its amount is noted as $-SiO_2$ (negative silica).

Molecular compositions of pyroxenes in the Morotu rocks calculated by this method are given in Table 17.

TABLE 17.—MOLECULAR COMPOSITION OF PYROXENES

	1305K*	1507*	607*	1305G*	902*	8203*	1412K*	1411*	1412S*	1412H*
Ac	—	4.62	4.16	13.86	11.09	32.34	58.67	64.22	69.76	81.31
Jd	—	—	0.40	—	2.02	—	—	—	—	4.85
Hd	16.41	24.05	26.53	26.04	32.49	29.27	23.56	16.12	11.65	7.69
Di	82.62	62.64	55.30	52.70	42.34	15.55	3.24	4.54	1.73	0.22
Fs	0.24	1.19	2.51	3.04	4.49	6.86	6.86	6.47	5.28	1.45
En	0.73	4.10	4.60	5.40	7.80	3.10	0.90	1.60	0.60	—
SiO_2	—	—	—	−3.72	−1.86	—	0.48	1.50	5.70	1.92
TiO_2	—	—	—	—	—	2.00	1.19	2.34	0.72	0.68
Al_2O_3	—	1.94	4.59	—	—	4.76	2.64	2.36	3.06	1.63
Fe_2O_3	—	—	—	1.44	—	4.32	2.08	0.48	0.16	—
Ac + Jd	—	4.8	4.9	13.7	13.1	37.1	62.9	69.1	78.4	90.2
Hd + Fs	16.6	26.1	31.0	28.8	36.9	41.5	32.6	24.3	19.0	9.6
Di + En	83.4	69.1	64.1	57.5	50.0	21.4	4.5	6.6	2.6	0.2
Wo	52.0	48.8	47.3	46.4	43.6	40.2	36.9	34.8	33.1	39.7
En	39.0	36.0	34.0	34.2	31.4	18.8	7.0	12.8	7.3	1.1
Fs	9.0	15.2	18.7	19.4	25.0	41.0	56.1	52.4	59.6	59.2

* See specimen descriptions under numbers in text.

From the manner of zonal structure in the crystals as well as from their paragenetic relations with other minerals, the trend of crystallization of pyroxenes in the Morotu rocks is determined as follows: diopsidic augite → titanaugite → titaniferous soda-augite → soda-augite → aegirinaugite → aegirine.

A molecular variation diagram is constructed, taking the above-mentioned order of crystallization as the abscissa and the contents of each molecule as the ordinate (Fig. 11). The composition of the diopsidic augite (No. 1305K) of the aggregate in the dolerite is estimated from the optical properties as well as the bulk composition of the aggregate as Wo_{52} En_{39} Fs_9. Di molecule decreases rapidly; Ac molecule, on the other hand, increases remarkably toward the later stage, while Hd molecule has a weak maximum in the soda-augite, or in the middle course of the evolution. En, Fs, and Jd molecules are all low throughout the course and show no pronounced variation.

In the Wo-En-Fs diagram (Fig. 12), all pyroxenes lie below the Di-Hd line, and the trend is nearly parallel to the conjugation line

Di-Hd, beginning from earlier Di-rich members to later Hd-rich ones. Since, in this diagram, the content of Ac molecule, which plays an important role in the later stage, is not included, complete solid solution in the system Di-Hd-Ac, and he demonstrated the trend of pyroxenes and the course of crystallization differentiation within this ternary system by calculation, start-

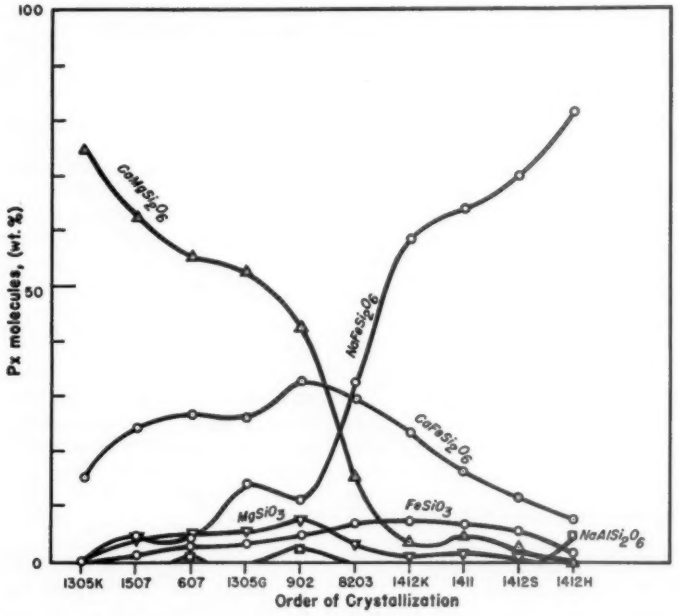

FIGURE 11. VARIATION DIAGRAM OF THE PYROXENE MOLECULES IN THE CRYSTALLIZATION SERIES OF THE MOROTU ROCKS

the diagram cannot provide a true picture of the evolution of aegirinaugite and aegirine. The whole trend is clearly given by the triangular diagram (Di + En)-(Hd + Fs)-(Ac + Jd) as shown in Figure 13. All pyroxenes lie on a smooth curve from diopsidic augite No. 1305K to nearly pure aegirine No. 1412H. This smooth and continuous course suggests that there may be continuous solid solution between these components.

Although there is not yet an experimental study on the Di-Hd-Ac system as a whole, it is suggestive that Schairer (unpublished data) has found that acmite and diopside form a complete solid solution of plagioclase type, disregarding the complication near the acmite end, introduced by its incongruent melting phenomenon. From his statistical study Vogt (1924, p. 112) concluded that there should be a ing with a composition of $Di_{54}Hd_{30}Ac_{16}$. It is interesting that this trend coincides perfectly with the one established in the Morotu pyroxenes. It is expected, therefore, that diopside, hedenbergite, and acmite may form complete solid solutions with some complications caused by the incongruent melting of acmite and by the transformation of hedenbergite into a wollastonite solid solution (Bowen, Schairer, and Posnjak, 1933, p. 217).

Tomita (1937, p. 976) gave three trends of crystallization of pyroxenes in the alkalic rocks as follows:

1. Augite → soda-augite → aegirinaugite → aegirine

2. Soda-diopside → soda-augite → aegirinaugite → aegirine

3. Titanaugite → aegirinaugite → aegirine.

Usually sodic pryoxenes do not crystallize out

until a certain stage is attained. The first trend is the most common and corresponds to that of the Morotu pyroxenes.

Recently the course of crystallization of pyroxenes from basaltic magma has been discussed

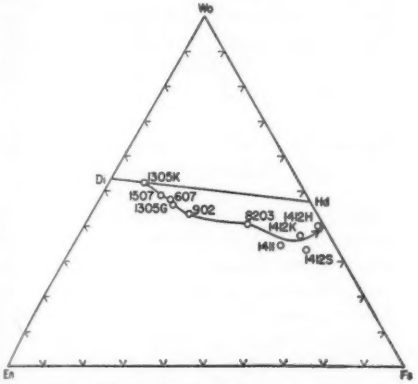

FIGURE 12. Wo-En-Fs DIAGRAM OF THE PYROXENES OF THE MOROTU ROCKS
Weight per cent

in detail by petrologists (Hess, 1941, p. 584–591; Edwards, 1942, p. 593; Kuno, 1950, p. 996). Compared with the course of crystallization of pyroxenes in calcalkalic rocks, the trend in the alkalic rocks is much simpler. This difference is made quite clear by the inspection of the Wo-En-Fs diagrams of the pyroxenes from the two rock series. Neither rhombic pyroxene nor pigeonite crystallizes out in the alkalic rocks, and the monoclinic pyroxene, which is the sole pyroxene phase, proceeds at first from diopsidic augite toward hedenbergite and then veers toward aegirine.

Augites from alkalic rocks are usually richer in Wo molecule than augites from calcalkalic rocks. This is partly a result of the crystallization of a large amount of olivine in the alkalic rocks which uses up much iron and magnesium silicates, leaving relatively more Wo molecule in the pyroxene components. The pyroxenes always fall near the Di-Hd join, but never within the Di-Hd-Wo triangle, as Tomita remarked (1937, p. 953).

When the molecular composition of pyroxenes is compared with the normative composition of their parent rocks, we can observe an intimate relation between the kind of pyroxene and the type of rock. Titanaugite or soda-augite from the dolerites or monzonites with normative nepheline and/or olivine is deficient in silica or shows a small amount of remainder components, if any. On the other hand, aegirinaguite or

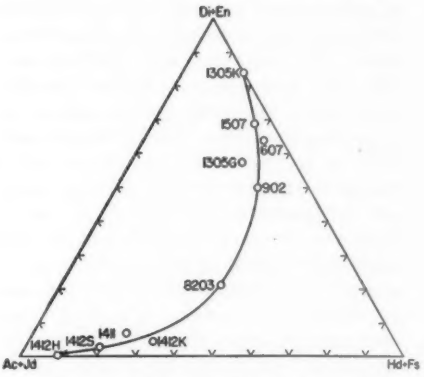

FIGURE 13. CRYSTALLIZATION COURSE OF THE PYROXENES OF THE MOROTU ROCKS IN A Di + En—Hd + Fs—Ac + Jd DIAGRAM
Weight per cent

aegirine from syenites without normative nepheline and olivine has excess SiO_2 or a high amount of remainder components. If the normative compositions are calculated for the pyroxenes just as for the rocks, the former pyroxenes always have normative nepheline, and the latter pyroxenes always have normative quartz. A similar relation was found by Angel (1929, p. 459), who showed that aerigine from granites has excess silica calculated as normative quartz; whereas aegirine from nepheline syenites or foyaites has normative nepheline. However, aegirine in the Morotu rocks is saturated. Thus the presence of normative nepheline in the present rocks should be ascribed not only to analcite, as previously remarked, but also to the chemical compositions of the pyroxene.

AMPHIBOLE: The amphiboles in the Morotu rocks include kaersutite, barkevikite, hastingsite, and arfvedsonite. The evolution of amphiboles is, however, not as clear as that of pyroxenes, because the zonal structure so common in pyroxenes is not prevalent in the amphiboles. Therefore, the paragenetic relations with other minerals is the most important means of de-

ducing the trend of crystallization of amphiboles.

Barkevikite is confined to the dolerites and is associated with basic plagioclase. It is rarely, if ever, present in monzonites or syenites. Kaersutite is abundant in the monzonites and also in some syenites, but a direct relation between barkevikite and kaersutite is not observed. Kaersutite is sometimes surrounded by greenish-brown hastingsite, with gradual transition, suggesting a continuous reaction relation between the two kinds of amphiboles. However, it is not uncommon to find a narrow belt of titanbiotite present between the two amphiboles. Arfvedsonite occurs mainly in the syenites and also in some monzonites in small amount. Though it rarely forms a fringe around greenish hastingsite, zonal growth with other amphiboles is not common. It is closely associated with aegirine of latest crystallization, sometimes exhibiting remarkable parallel growth. It is therefore evident that arfvedsonite is the last amphibole to crystallize. From these paragenetic relations the trend of amphiboles is determined as follows:

Barkevikite → kaersutite → hastingsite → arfvedsonite.

There are insufficient data to decide whether or not this trend is continuous throughout the series. Though the variation in chemical composition throughout the amphibole trend is not definitely demonstrated in the diagram, the comparison of the compositions of kaersutite, an earlier member, and arfvedsonite, a later member, shows that alkalis and iron oxides increase and MgO and CaO decrease toward the later stages of the trend, while TiO_2 reaches its maximum in the middle course. Aenigmatite shows no genetic association with other amphiboles, and its place in this trend cannot be settled.

BIOTITES: Though there is some difference in the chemical composition of the titanbiotite from the dolerite and monzonite, it is not so great as the difference between the parent rocks. Though lepidomelanic biotite, which is confined to the syenites, was not chemically analyzed, it is very rich in iron and rather poor in titanium, judging from its deep brown color (Hall, 1941, p. 32). There is no association of titanbiotite and lepidomelanic biotite. These observations show that the titanbiotites remained fairly constant in their chemical composition throughout the crystallization, disappearing before the stage of the crystallization of syenites was attained, and that lepidomelanic biotite crystallized out in the latest stage. There is no obvious indication whether or not the relation between titanbiotite and lepidomelanic biotite is continuous.

Paragenetic Relation of Aenigmatite, Aegirinaugite, Arfvedsonite, and Aegirine

Zonal structure proceeding from a core of aenigmatite through aegirinaugite and arfvedsonite to the margin of aegirine is one of the most interesting paragenetic relations in the mafic minerals. The chemical compositions of the four minerals are given in Table 18. An average composition of 14 samples of aenigmatite from the various localities of the world is shown in column 1, for no chemical analysis was made on the aenigmatite in the Morotu rocks. Aenigmatite is characterized by very high FeO and TiO_2, rather high Na_2O, and very low Fe_2O_3. It is evident from Table 18, in which the minerals are arranged in the order of their crystallization, that FeO and TiO_2 rapidly decrease, Fe_2O_3 rapidly increases, and Na_2O moderately increases toward the later stage of crystallization. Aegirinaugite and arfvedsonite are much alike in their composition; disregarding H_2O+, the main difference is the proportion of FeO and Fe_2O_3, while the amount of total iron is nearly unchanged. Pyroxene molecules calculated for aenigmatite and arfvedsonite as well as for the actual pyroxenes, aegirinaugite and aegirine, are plotted in the (Di + En)-(Hd + Fs)-(Ac + Jd) diagram (Fig. 14). The evolution from aenigmatite to aegirine is represented by increase of soda silicates at the expense of iron silicates, while magnesium silicates always remain low. As aegirinaugite and arfvedsonite have somewhat similar composition, it is inferred that whether the one or the other will crystallize out will depend on the presence of volatile components, especially of water, which favors the crystallization of arfvedsonite. Aegirinaugite, which crystallizes out when the concentra-

TABLE 18.—CHEMICAL COMPOSITION OF SOME ALKALI MAFIC MINERALS IN A ZONAL STRUCTURE

	(1)	(2)	(3)	(4)
SiO_2	40.14	47.82	48.51	51.51
TiO_2	8.40	1.19	1.10	0.68
Al_2O_3	1.80	2.64	1.08	2.86
Fe_2O_3	5.32	22.36	17.53	28.12
FeO	33.17	10.58	14.31	3.01
MnO	1.96	0.00	0.66	0.00
MgO	1.10	0.94	1.31	0.04
CaO	1.15	6.14	5.43	1.80
Na_2O	6.01	7.78	7.87	11.48
K_2O	0.26	0.16	1.48	0.32
H_2O+	0.62	—	1.48	—
H_2O-	—	1.16	0.17	0.24
Total.....	99.93	100.77	100.93	100.06

1. Aenigmatite, average of 14 analyses of aenigmatites from various localities of the world
2. Aegirinaugite No. 1412K, Morotu
3. Arfvedsonite No. 1412, Morotu
4. Aegirine No. 1412H, Morotu

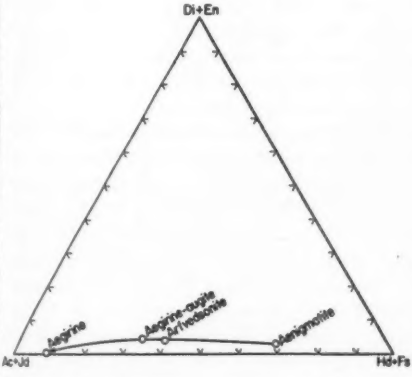

FIGURE 14. CRYSTALLIZATION COURSE OF AENIGMATITE-ARFVEDSONITE-AEGIRINAUGITE-AEGIRINE SERIES IN A Di + En—Hd + Fs—Ac + Jd DIAGRAM

Weight per cent

tion of water is lower, gradually becomes unstable, and its place is taken by arfvedsonite when the magma becomes more and more enriched in water as crystallization proceeds. However, arfvedsonite in turn ceases to be stable when the residual magma becomes very rich in soda, and aegirine which has the highest soda content will eventually crystallize out in spite of the presence of abundant water.

Vogt (1924, p. 119) suggested that the water content in the magma is the main factor in determining whether aegirine or soda-hornblende will crystallize out and that both minerals may form simultaneously only when the magma has a "medium" content of water. However, this suggestion does not seem applicable in this case, for aegirine crystallized out in the latest stage, when the water content is supposed to be highest. Therefore, it seems to be more reasonable to ascribe the crystallization of aegirine to the predominance of soda in the residual solutions. Tomita (1938, p. 736) offered a similar view on the crystallization of aegirine and some alkalic hornblende.

Reaction Relation of Iron Ores with Other Mafic Minerals

In the dolerites, titanbiotite is often intergrown with the ores, and barkevikite is formed as a reaction rim around the ores enclosed in the titanaugites. In some monzonites and syenites, soda-augite gives way to aegirine along the boundary between it and the ore. Arfvedsonite often mantles iron ores in the syenites.

Bowen, Schairer, and Willems (1930, p. 427) have shown that acmite melts incongruently at 990° C. according to the following reaction:

$$Na_2O \cdot Fe_2O_3 \cdot 4SiO_2 = Fe_2O_3 + \text{soda-rich solution.}$$

Therefore, the reaction of soda-rich residual solution on the hematite, whose presence in these rocks has been proved, produced at least some of the aegirine, although the greater part may have crystallized directly from the magma. Though neither arfvedsonite nor barkevikite has been studied experimentally, their similar behavior in the rocks concerned suggests that some of these hornblendes might also be formed in a similar manner. Such a reaction relation is absent between olivine and ores, even where the two minerals occur side by side. The reaction relation of iron ores is always confined to the alkalic mafic minerals of relatively later stage. The important role played by titaniferous iron ores in the alkalic rocks in general, there-

fore, may be ascribed to the effect of highly active soda-rich solutions on these ores in the later stage. A reaction relation between iron ore and mafic minerals is usually not so common in the calcalkalic rocks. However, similar reaction is observed in cases when sodic solutions might have concentrated, as in some granitic rocks.

Stability Relation between Biotite and Aegirine

One notable feature in the mineral association is that titanbiotite, an essential component in dolerites and in some monzonites, is either absent or is replaced by lepidomelanic biotite in trifling amount in the syenites. A similar relationship is frequently observed in alkalic rocks from many other localities.

If the parental magma is rather rich in potash and water, biotite will probably crystallize out in place of orthoclase, leucite, and olivine. In this connection the following observations on the genetic relation between leucite and biotite deserve special attention. Washington (1907, p. 386) suggested that leucite instead of biotite is formed when the pressure of water vapor is lowered. Thus in Highwood Peak, Montana, biotite-rich rock often passes into a leucite-bearing rock within a single dike where the escape of water was facilitated (Pirsson, 1905). A similar relation was observed by Ogura and Matsuda (1936, p. 60) in a single lava flow of the so-called "shihlunite" (leucite basalt) of Wutalienchih Volcano, Manchuria. There the quenched surface is leucite-bearing glassy rocks, and the inner holocrystalline part is rich in biotite.

On the other hand, the biotite molecule ceases to be stable in the residual magma enriched in soda and ferric iron, and aegirine as well as orthoclase will separate instead. Water will form some hydrous minerals, such as arfvedsonite or analcite, in the deuteric stage. This reaction is in agreement with the continual enrichment of the feldspar in Or molecule with increase of aegirine, as is actually observed in the syenites. That a large amount of silica is necessary for the formation of aegirine in place of biotite explains the absence of quartz even in the syenites rather rich in silica.

Petrogenesis

Parental Magma

The predominance of dolerites among the Morotu rocks and their similarity in mineralogical as well as chemical composition justify use of the average composition of the dolerites for the discussion of the probable parental magma. This average is given in column 1 of Table 19, together with the composition of basaltic rocks of the world for comparison. The Morotu average is very similar to the average composition of olivine basalts of the Eastern Asiatic alkalic province as given by Tomita (1935, p. 294), while it differs more or less from that of any of the tholeiitic basalts given in columns 3–7. A normative pyroxene diagram (Fig. 15) shows this relation more clearly. Normative pyroxenes of Morotu dolerites and Asiatic basalts are rich in Wo molecule, falling near the Di-Hd line; on the other hand, those of tholeiitic basalts are poor in Wo and cluster in the central part of the Di-Hd-Fs-En field. It is interesting that Macdonald's average of Hawaiian "olivine basalts," which, according to Tilley (1950, p. 40), include both olivine basalts and tholeiites, lies between the two groups mentioned above. From these considerations, it is clear that the Morotu dolerites belong to the olivine basalt magma.

The genetic relation between tholeiitic and olivine basalt magma has frequently been discussed (Kennedy, 1933, p. 239; Kennedy and Andersen, 1938, p. 38; Tomita, 1935, p. 300; Barth, 1936, p. 350; Barth, Correns, and Eskola, 1939, p. 80). Kuno (1950, p. 1012), from study of the volcanic rocks of Hakone volcano, concluded that the parental position of olivine basalt magma to the tholeiitic magma is justified. Tilley (1950, p. 45), on the other hand, drawing attention to the frequent occurrence of tholeiites in Hawaiian or other oceanic islands, proposed that tholeiitic magma is the original and that crystallization differentiation is responsible for further derivation of two rock series, alkalic and calcalkalic, from the tholeiitic magma.

In this connection, it is interesting that in the Morotu district there are many Miocene volcanic rocks such as andesite, dacite, and their pyroclastics, all of which are calcalkalic.

Although it is apparent that these rocks are older than the Morotu alkalic rocks, mutual relation of these two series could not be found in the field. It is not known whether these two rock series were independently derived from different parental magmas or were formed from a common magma.

However, it may be concluded that the alkalic rocks of the Morotu district were derived from olivine basaltic magma with moderately alkalic affinity, no matter how the magma was formed.

Table 19.—Chemical Composition Of The Basaltic Rocks Of Various Petrographic Provinces Of The World

	(1)	(2)	(3)	(4)	(5)	(6)	(7)
SiO_2	46.50	49.05	48.34	48.80	52.3	48.35	52.5
TiO_2	2.25	2.28	1.22	2.19	1.1	2.77	1.0
Al_2O_3	16.42	16.48	18.80	13.98	14.5	13.18	15.4
Fe_2O_3	4.05	4.00	3.67	3.59	3.8	2.35	1.2
FeO	6.73	6.80	6.40	9.78	10.2	9.08	9.3
MnO	0.12	0.23	0.19	0.17	0.3	0.14	0.2
MgO	6.35	5.60	5.13	6.70	5.1	9.72	7.1
CaO	8.86	9.31	10.22	9.38	10.0	10.34	10.3
Na_2O	3.10	35.7	2.66	2.59	2.1	2.42	2.1
K_2O	1.54	1.99	0.94	0.69	0.4	0.58	0.8
H_2O+	3.40	—}	2.28	1.80	—	—	—
H_2O-	0.43	—}		—	—	—	—
P_2O_5	0.25	0.69	0.38	0.33	0.2	0.34	0.1
Total.....	100.00	100.00	100.00	100.00	100.0	99.27	100.0
Or	15	18	9	7	5	7	9
Ab	42	45	35	44	36	43	34
An	43	37	56	49	59	50	57
Wo	49	52	20	23	24	33	21
En	39	34	52	46	36	47	43
Fs	12	14	28	31	40	20	36

1. Average of 4 dolerites of the Morotu district
2. Average of 50 basalts of Eastern Asiatic alkalic province (Tomita, 1935, p. 294)
3. Average of 36 basaltic rocks of Japan, excluding alkalic rocks (Iwasaki, 1937, p. 351)
4. Average of 19 plateau basalts of the world (Daly, 1933, p. 17)
5. Average of the groundmass composition of 11 basalts of the Izu district, Japan (Kuno, 1950, p. 1009)
6. Average of 53 olivine basalts of the Hawaiian Archipelago (Macdonald, 1949, p. 1569)
7. Average of 43 dolerites of Karroo region, Africa (Walker and Poldervaart, 1949, p. 649)

Enrichment of Iron in Basaltic Magma

Chemical composition of the Morotu rocks is plotted in the total FeO-MgO-Alkalic diagram (Fig. 16), together with the Skaergaard liquids (Wager and Deer, 1939, p. 314), Daly's averages for basalts, andesites, dacites, and rhyolites (Daly, 1933, p. 9), and alkalic rocks of the Mull and Ardnamurchan districts (Thomas and Baily, 1924, p. 13). Absolute enrichment in iron is not observed throughout the course of crystallization of the Morotu rocks, in spite of the fairly high iron content compared with the very small amount of MgO in the syenites. The Morotu rocks fall entirely within the area occupied by the Mull and Ardnamurchan rocks.

Relative enrichment of iron compared to MgO in basaltic magma during crystallization has long been recognized. Wager and Deer have shown that there was an absolute enrichment in iron during the early stage of crystallization of strongly fractionated tholeiitic magma of

Skaergaard, and this has also been observed in basaltic rocks from other parts of the world

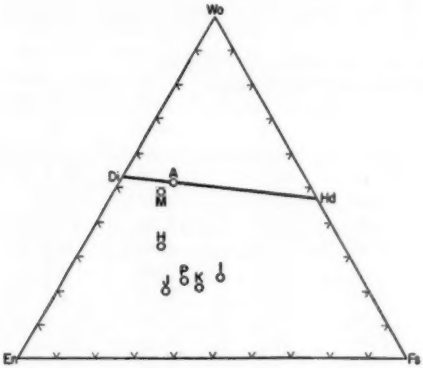

FIGURE 15. NORMATIVE PYROXENE OF BASALTIC ROCKS

M.....Morotu
A.....Eastern Asia
H.....Hawaii
I.....Izu
J.....Japan
K.....Karroo
P.....Plateau Basalts

Weight per cent

(Walker and Poldervaart, 1949, p. 656; Edwards, 1942, p. 603).

The Morotu magma was of such a composition that large amounts of biotite or amphibole with high iron content separated out early, and

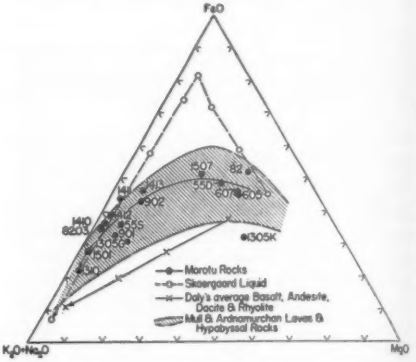

FIGURE 16. COMPOSITION OF THE MOROTU ROCKS PLOTTED IN A $K_2O + Na_2O$—total FeO-MgO DIAGRAM

not much iron was left in the residual liquids. On the other hand, fractionation of feldspars was predominant and controlled the trend toward enrichment in alkalis and silica. So far the known cases of absolute iron enrichment are

PLATE 1.—PHOTOMICROGRAPHS OF THE MOROTU ROCKS

FIGURE 1.—DOLERITE NO. 601

Showing olivine phenocryst rimmed with alteration products in coarse-grained groundmass composed of titanaugite and labradorite. × 31.

FIGURE 2.—DOLERITE NO. 605

Showing well-developed ophitic texture of titanaugite and labradorite. Black minerals are either titanbiotite or titaniferous iron ore. × 7.

FIGURE 3.—DOLERITE NO. 1507

Showing intergrowth of titanaugite and titaniferous iron ore, which is also surrounded by thin rim of titanbiotite. Analcite is seen at the upper left of titanaugite. × 7.

FIGURE 4.—DOLERITE NO. 55D

Showing intergrowth of titanaugite and titaniferous iron ore. Titanaugite becomes hastingsite or titanbiotite in those parts in contact with iron ore. × 31.

PLATE 2 —PHOTOMICROGRAPHS OF THE MOROTU ROCKS

FIGURE 1 —MONZONITE NO. 1219

Zonal structure in feldspar. Plagioclase (Pl) in the core in part changed to analcite (gray patches). Turbid microperthite (Mp) makes greater part of the mantle, and clear anorthoclase (An) occurs sporadically as irregular patches. Outermost thin rim is nearly pure and fresh albite. × 31.

FIGURE 2.—MONZONITE NO. 901

Showing similar zonal structure of feldspar as in Figure 1. In the lower right druse, analcite (Ac) is developed. × 31.

FIGURE 3.—SYENITE NO. 1410

Similar zonal structure of feldspar. Clear, large crystals of albite (Ab) are developed in the druse. × 31.

FIGURE 4.—SYENITE NO. 1412

Microperthite makes the bulk of the feldspars. Small euhedral plagioclase with thin albite rim is in the center. Crossed nicols. × 31.

FIGURE 2

FIGURE 4

FIGURE 1

FIGURE 3

PHOTOMICROGRAPHS OF THE MOROTU ROCKS

FIGURE 1

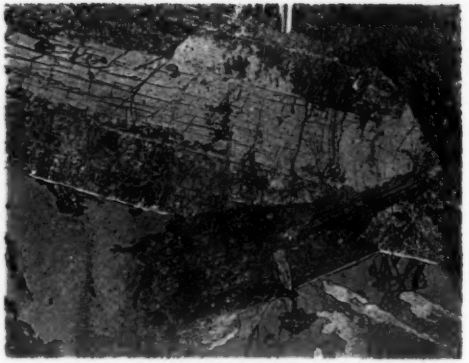

FIGURE 2

FIGURE 3

FIGURE 4

PHOTOMICROGRAPHS OF THE MOROTU ROCKS

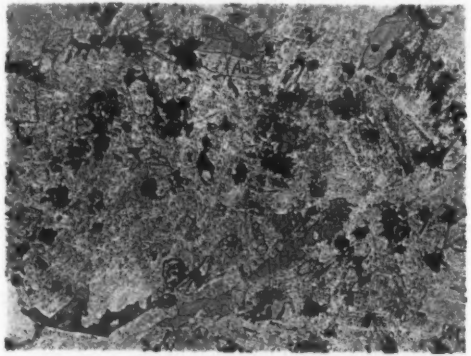

FIGURE 1

FIGURE 2

FIGURE 3

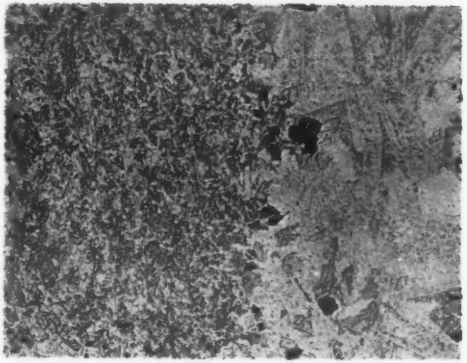

FIGURE 4

PHOTOMICROGRAPHS OF THE MOROTU ROCKS

PHOTOMICROGRAPHS OF THE MOROTU ROCKS

FIGURE 2

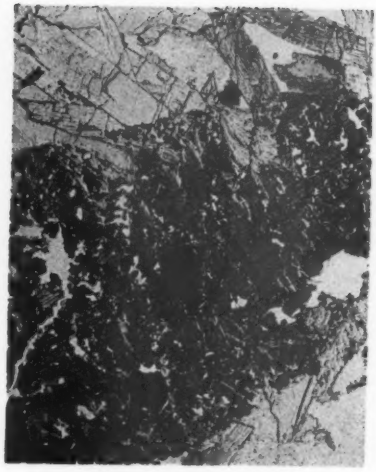

FIGURE 4

FIGURE 1

FIGURE 3

restricted to calcalkalic rocks. Edwards remarked that iron enrichment occurs in closed chambers such as sheets or laccoliths, but does not occur in the subjacent bodies without bottoms, and he concluded that shape of the intrusive body is the controlling factor.

Iron enrichment did not occur, however, in the Morotu rocks, although they commonly form laccoliths or sheets. Therefore, it is clear from our observation that not only the shape of the bodies, but also composition of the magma, decides whether iron enrichment or alkali and silica enrichment will prevail in the intrusive bodies.

Crystallization Differentiation

A prominent feature of the Morotu rocks is the close association of monzonite and syenite with dolerite in a single intrusive body. This association is fairly common among the oceanic islands, where trachyte is accompanied by olivine basalt (Daly, 1927, p. 70; Esenwein, 1929, p. 128). Closer similarity is found, however, in the classical example of Shonkin Sag laccolith (Weed and Pirsson, 1901, p. 11; Pirsson, 1905, p. 1; Barksdale, 1937, p. 354; Hurlbut and Griggs, 1939, p. 1043), or in the sills of Ayrshire, Scotland (Tyrrell, 1928, p. 540). Recently, similarly differentiated sheets were found by the writer in the Nemuro Peninsula, Hokkaido, Japan (Yagi, 1948, p. 323).

The following points are common to all these occurrences:

(1) The main parts of the dikes or sheets are always mafic rocks such as dolerite or basalt.

(2) Felsic parts such as syenite or monzonite are never found in direct contact with the country rocks, but are confined to the central parts of the bodies or form irregular schlieren or veins only within mafic parts.

(3) Boundaries between mafic and felsic parts

PLATE 3.—PHOTOMICROGRAPHS OF THE MOROTU ROCKS

FIGURE 1.—MONZONITE No. 1413

Hastingsite (gray) and soda-augite (light gray) in parallel growth. Biotite (dark gray) is usually associated with iron ore (black). Matrix is plagioclase and microperthite. × 29.

FIGURE 2.—MONZONITE No. 84

Showing parallel growth of kaersutite (Ka), titanbiotite (Bi), and soda-augite (Au). Biotite is developed along the boundary of the other two minerals. Note skeletal crystal form of iron ore. Long needle crystal of apatite in the upper left corner. × 27.

FIGURE 3.—MONZONITE No. 1305G

Titaniferous soda-augite (Au), titanbiotite (Bi), and kaersutite in parallel intergrowth. Augite on the lower left corner shows hour-glass structure. Kaersutite in the upper part has numerous minute inclusions arranged at 45° to cleavages. × 24.

FIGURE 4.—SYENITE No. 55S AND DOLERITE No. 55D

Syenite (right half) has albite as main mineral with interstitial soda-augite and iron ore. Dolerite (left half) shows ophitic texture composed of fine grains of barkevikite (dark gray), titanaugite (light gray), plagioclase, anorthoclase, and iron ore (black). × 31.

PLATE 4.—PHOTOMICROGRAPHS OF THE MOROTU ROCKS

FIGURE 1.—SYENITE No. 1501

Aegirine (gray) and slightly turbid microperthite (slightly grayish) are the main components. Analcite (white) is developed interstitially between microperthites. × 31.

FIGURE 2.—SYENITE No. 1411

Showing parallel growth of aenigmatite and rhönite (black) in the center and in the right side. Long acicular crystals of aegirine form aggregates radiating from aenigmatite. Matrix is composed mainly of microperthite. × 7.

FIGURE 3.—SYENITE No. 1411

Large arfvedsonite crystal (Ar) forms a parallel growth with small aegirine (Ae). Parallel growth of aenigmatite (dark gray) and rhönite (black) is shown in the right side. × 27.

FIGURE 4.—SYENITE No. 1411

Large crystal of aenigmatite (dark gray) includes irregular patches of rhönite (black) in parallel growth. On the right, aegirine (light gray) forms prismatic crystals and, rarely, radial fibrous crystals. × 29.

are either transitional or sharp; in neither case, however, are chilled margins found in the felsic parts.

When the parental olivine basaltic magma was intruded into the present horizon, crystal-

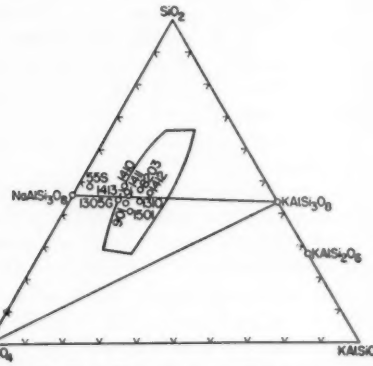

FIGURE 17. LOW-TEMPERATURE VALLEY OF THE SYSTEM $KAlSiO_4$—$NaAlSiO_4$—SiO_2, WITH PROJECTION OF MONZONITES AND SYENITES OF THE MOROTU DISTRICT

Weight per cent

lization began from the cooler part, forming dolerites in the marginal parts; consequently, the residual liquids in the inner parts always showed an increasing proportion of alkalis, silica, and volatile components. When Morotu syenites and monzonites with more than 70 per cent normative salic minerals are plotted in the $NaAlSiO_4$-$KAlSiO_4$-SiO_2 diagram (Fig. 17), all but one of the 10 projections lie within the lowest trough of "petrogeny's residual system" (Bowen, 1937, p. 11). This provides strong evidence that these felsic facies formed from the residual liquid under the dominant control of fractional crystallization, but the mechanism controlling the fractionation is not known. Gravitational separation of early-formed crystals may have been responsible, but mafic layers of accumulated crystals have not been observed in the exposed outcrops in this district. The continuing concentration of volatiles in the residue as crystallization proceeded was no doubt a major factor.

Syenite No. 55S, the only exception which lies outside this trough, is very rich in Ab, and this rock, as well as soda-rich dolerite No. 55D, constitutes the second minor trend of change in the normative feldspar diagram. Soda enrichment by some metasomatic process may be responsible for this, but definite evidence is lacking.

The syenitic or monzonitic residual liquids, highly charged with volatiles, were very mobile and could migrate by their own vapor pressure or by the hydrostatic pressure of superincumbent rocks into fissures in the already solidified dolerites. In such case the boundary between these veins and the dolerites is sharp, though there is no chilled margin in the veins. Where the residual liquids gradually solidified *in situ*, the boundaries are transitional.

Abundance of analcite is worthy of notice. Yoder (1950, p. 323) has shown experimentally that analcite could be in equilibrium directly with the silicate liquids at pressures higher than, say, 10,000 atmospheres. No such condition could have been realized in the present case. Friedman's experiments (1951, p. 26) confirmed that analcite can crystallize out directly from aqueous solution containing soda, alumina, and silica. Analcite in the Morotu rocks formed in the hydrothermal stage in analogous manner. The occurrence of aegirine, aegirinaugite, and arfvedsonite crystals in miarolitic cavities is further clear evidence of hydrothermal crystallization. This is also supported by the experiments by Königsberger and Müller (1906, p. 339).

Therefore, it is clear that abundant volatile components in the Morotu magma played an essential role in the later stage, especially during the formation of syenites or monzonites.

Morotu District as a Petrographic Province

There are several localities of alkalic rocks in Japan. Oki Island in the Japan Sea, among others studied by Kozu (1913, p. 25) and Tomita (1927–1932), exhibits a most interesting suite. Mainly effusive with minor intrusive dikes, these alkalic rocks cover a wide range from trachyandesitic basalt to comendite. No feldspathoidal rocks are found. When compared with the Morotu rocks, the presence of hypersthene, pigeonite, riebeckite, quartz, etc., and the absence of biotite, barkevikite, analcite, microperthite, etc., in the Oki rocks are the striking differences in the mineral association. Other alkalic volcanic rocks are trachyte and

alkalic basalt of northern Kyusyu (Kozu, 1911, p. 555, 566); melilite-bearing nepheline basalt of Nagahama, Simane Prefecture; volcanic and hypabyssal rocks comprising basalt, teschenite, mugearite, or trachyte of Abekawa volcanic zone, Sizuoka Prefecture (Yamazaki, 1930, p. 435); and trachyte of Atumi, Yamagata Prefecture (Nakano and Kawano, 1936, p. 156). They are all located along the Japan Sea, with the exception of the Abekawa zone, which is near the Pacific Ocean.

Many localities of alkalic rocks are known in Korea. Hakuto Volcano (Kozu and Seto, 1928, p. 781; Watanabe, 1934, p. 40) and Kainei district (Ichimura, 1931, p. 215), both on the Manchurian border, and Meisen district (Yamanari, 1925, p. 1; Tateiwa, 1925, p. 1) in eastern Korea have various kinds of volcanic rocks ranging from basalts to comendites. Some of them are similar to the Morotu rocks in modes of occurrence as well as in mineral association. Saisyu Island in the Yellow Sea between Japan and Korea (Haraguchi, 1930, p. 595) has rocks ranging from basalt to trachyte. The petrochemical character of this alkalic suite is quite similar to that of Morotu rocks.

Uturyo Island off the Korean coast in the Japan Sea is unique for its leucite-bearing rocks, such as vicoite (Tsuboi, 1920, p. 91). Wutalienchih Volcano in Manchuria is another locality for leucite, and Ogura and Matsuda (1936, p. 1) gave the name shihlunite to a peculiar leucite basalt from this volcano. Volcanic rocks of both localities are richer in potash than in soda.

All these localities lie within or along the periphery of the Japan Sea, and their alkalic rocks, which are mostly effusive with smaller amount of intrusive, are products of the igneous activity in the Cenozoic Era. Thus they constitute a well-defined Circum-Japan Sea province of Cenozoic alkalic rocks (Tomita, 1935, p. 227).

From its geographical situation, its period of activity, as well as from its petrographic character, it is evident that the Morotu district also belongs to this province and probably constitutes the northeastern boundary of this large petrographic area.

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STRUCTURE OF GRAYWACKE AREAS AND TACONIC RANGE, EAST OF TROY, NEW YORK

By Robert Balk

Abstract

The report deals with the structure of graywacke areas, east of the Hudson River near the latitude of Troy, New York, and the structure of a part of the Taconic Range in the same latitudes. It has been claimed that much of this area is underlain by allochthonous far-travelled masses of rocks, resting on flat thrusts.

The contact zones of the Rensselaer graywacke, and of smaller graywacke areas south of it, show that the graywacke layers are conformably interbedded with greenish and purple slates and siltstones which underlie large portions of the Hudson lowland. If thrusts are present, they must lie below the levels previously assumed, to take cognizance of the uninterrupted sequences of strata below the present edges of the graywacke terranes.

The coarsest phases of graywacke crop out near the western edges of the area. Here are also found the thickest sections of graywacke and several lenses of coarse, bouldery conglomerate. Eastward, fine-grained graywacke forms individual lenses in phyllitic slates and schists of the Taconic Range.

The Taconic Range consists of a thick mass of quartz-chlorite schist, with subordinate zones of gray siltstone, quartzitic and graphitic lenses, and a tuffaceous series of rocks along its western border. The rocks have been folded, overturned to the west, but isoclinal folds appear to be rare. Slip cleavage dips east, and east of the graywacke areas a well-developed lineation plunges east-southeast, parallel to the direction of movement along many thrust faults.

About 8 miles south of Bennington, Vermont, the rocks of the Taconic Range approach the Green Mountain front. This tract, referred to as the Pownal upland, is structurally complex, and unfortunately poor exposures leave many of its problems open. Flat thrusts may occur here, but the magnitude of their displacements is uncertain.

The structural relations between the Pownal upland and the Green Mountain front are discussed, and the writer attempts to reconstruct the tectonic history of the area in the light of those features of deformation that are exposed and believed to allow a fairly reliable interpretation. Folding, flowage, and thrust faulting of an autochthonous series of argillaceous Lower Cambrian and Ordovician rocks are thought to explain satisfactorily most, if not all, observable features of the region.

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INTRODUCTION

The "Taconic controversy" is famous in the history of North American geology. Although one particular problem, the age of rocks containing olenellid trilobites, has been settled, the structural problems of the Taconic Range in eartern New York still await solution. This paper aims to provide some facts and figures concerning the lithologic characteristics and structure of the Taconic area, east of Troy, New York. Where it has been necessary to add interpretations, this has been stated, and an effort has been made to give accurate locations of important outcrops so that others may find them.

The principal structural problem of the Taconic Range is the provenance and structure of the argillaceous rocks that compose this range, and the relationship of these rocks to surrounding formations of different lithology. In a number of brief announcements, Arthur Keith (1912; 1913) stated his belief that the argillaceous rocks at the extreme north end of the range, southwest of Brandon, Vermont, rest on foreign ground. Later (Keith, 1932, p. 364–366), he expanded this view to include appreciable parts of the range.

A portion of the Taconic Range, about 60 miles south of the Brandon area, was studied by L. M. Prindle who, with E. B. Knopf

reached the tentative conclusion (Prindle and Knopf, 1932) that not only the Taconic Range, but also the large Rensselaer graywacke upland, west of the range, may be remnants of thrust sheets.

Another 60 miles south, the writer studied a district near the southern end of the Taconic Range but found no evidence for far-travelled low-angle thrusts (Balk, 1936). In the meantime, the concept of Taconic thrusts has so stimulated geologic thinking that the authors of several modern compilations of paleogeography and structure of western New England favor the view that the entire range, 150 miles long and 15 miles in average width, as well as considerable areas near by, rest on foreign ground. It seems appropriate, therefore, to re-examine some of the facts upon which the assumptions of large-scale horizontal movements in the Taconic area are based.

The writer's interest in the region east of Troy was awakened by T. N. Dale's description of the "Rensselaer grit", one of the very few first-cycle graywackes in the Appalachian region (Dale, 1893). If its entire border, with a total length of nearly 50 miles, were the trace of a thrust, as considered by Prindle and Knopf, there seemed a good opportunity to study the Taconic thrust both inside and outside the range proper. A total of about 11 months of field work has been done intermittently since 1926. Rocks and minerals have been collected wherever lithologic or metamorphic questions seemed to be involved; 573 thin sections, and 7 chemical analyses have been made; and the fabric of several rocks has also been studied.

Scope of Work

Structure and lithologic characteristics of the rocks in this region are described from west to east. The largest mass of graywacke forms the so-called Rensselaer graywacke upland, about 10 miles east of Troy (Fig. 1). Its borders have been studied more carefully than the interior which is a till-covered, partly swampy plain with few outcrops, but, as no recent mapping in this region seems to have been done, some of the rocks within the upland are also described. South of the upland, there are several additional, smaller graywacke areas. If the large mass is the remnant of a thrust sheet, so are perhaps the smaller, adjoining graywacke districts. They have therefore also been examined, and their description follows that of the upland.

This is followed by a survey of the lowland, east of the graywacke upland, that narrows in the middle to the valley of Berlin, New York. The concluding areas are the Taconic Range proper, from Bennington, Vermont, on the north, to the vicinity of Hancock, Massachusetts; the valley east of that range, near the village of North Pownal, Vermont; and the unique upland that bridges the Taconic Range and the Green Mountains in Pownal township, Vermont. A few remarks concern the rocks of the Green Mountains where they adjoin the upland. A brief discussion of the nature of the deformation, with tentative reconstruction of the area of Cambro-Ordovician deposition, concludes the report.

Previous Work

Parts of this region have been mapped or referred to by other workers (Fig. 2). Additional papers are referred to in the text.

Acknowledgments

Field expenses in 1948 and 1949, as well as expenses for chemical analyses and a part of the thin sections, were met by the University of Chicago. The writer was privileged in being guided by Prindle to the important fossil locality at Moon Hill, and to other fossil localities north of the region; discussions with Prindle on the geology of the region have been most profitable. For a careful re-examination of fossils collected by C. D. Walcott, A. Foerste, and L. M. Prindle, and deposited at the United States National Museum, the writer is indebted to Christina Lochman. Her report on the fossils will appear elsewhere. The manuscript was reviewed by Marshall Kay and Ernst Cloos.

The Graywacke Areas

General Statement

From the latitude of Castleton, Vermont, southward to Austerlitz township, New York,

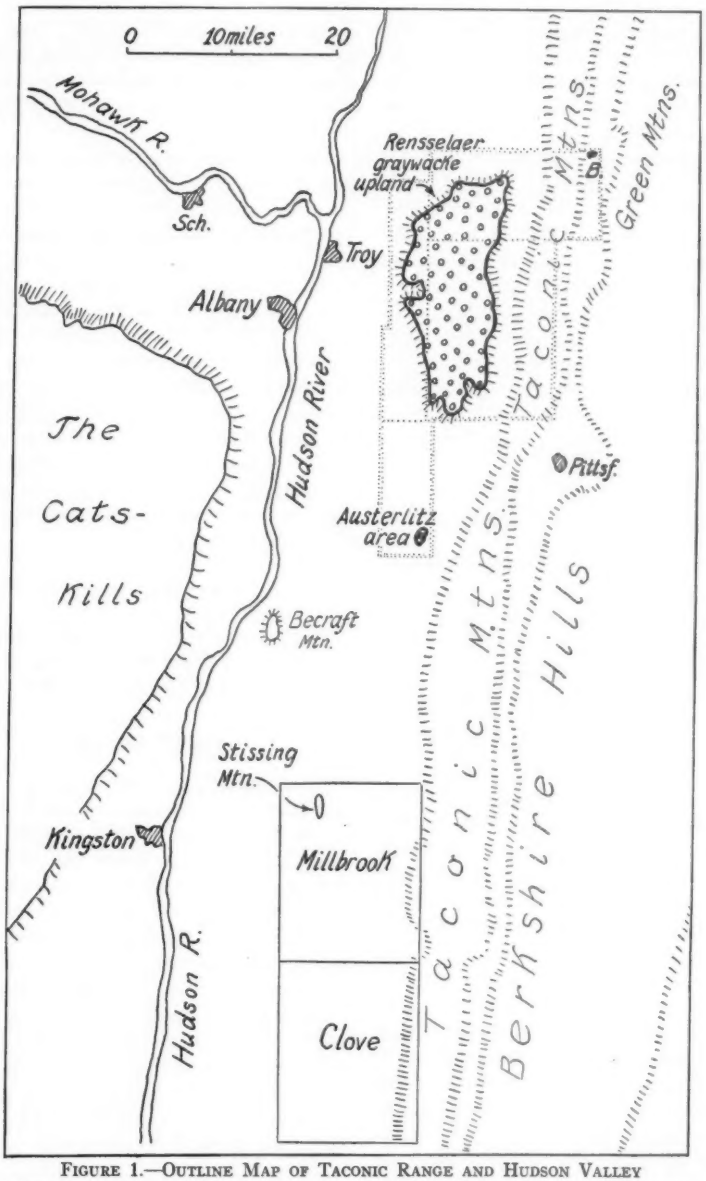

FIGURE 1.—OUTLINE MAP OF TACONIC RANGE AND HUDSON VALLEY

Area of Plate 1 indicated by double row of fine dots. B. = Bennington, Vt.; Sch. = Schenectady, N. Y.; Pittsf. = Pittsfield, Mass. Area covered by the 15′ quadrangles Millbrook and Clove is marked.

areas of graywacke appear on the west side of the Taconic Range over a distance of about 90 miles. The writer did not visit the northernmost area, at Bird Mountain, Vermont (15′ Castleton quadrangle, Vt.; Dale, 1893, p. 337-340; 1900; Kaiser, 1945, p. 1089–1090; Fowler,

1950, p. 40). Correlation and structure of all the graywacke areas have long been controversial subjects. Thus, T. Nelson Dale (1893, p. 330–331) concurred with W. W. Mather (1843, p. 368, 382, 384) and James Hall (1852, p. 1) in assigning the Rensselaer graywacke, east of Troy, to the Silurian, correlating it with the Shawangunk and Oneida conglomerate of the New York section. He expressed the same opinion in 1904 (Dale, 1904a, p. 42). Ruedemann (1930, p. 128) considered the rocks late Devonian, equivalent to the Catskill sandstone. Prindle and Knopf (1932, p. 284) assign the Rensselaer graywacke tentatively to the Lower Cambrian. Fowler (1950, p. 45) considers that the Bird Mountain graywacke may belong to the Nassau formation for which very early Cambrian, or even latest Precambrian, age has been proposed.

Dale (1893, p. 324–325) thought he could see a general synclinal structure in the Rensselaer graywacke. Ruedemann (1930, p. 151) was impressed with what he took to be recumbent and other types of intense folding in the graywacke, but thought the formation overlies pre-Devonian rocks unconformably. Prindle and Knopf (1932, Fig. 3) assume westward-overturned folds in the graywacke; but they believe it rests on a flat thrust plane. Despite these numerous contrasting views, one looks in vain for descriptions of the contacts of this remarkable rock. Presumably, it shows either an unconformity or a thrust at the base. To fill this gap, the writer has walked out the contacts of the graywacke areas and supplemented the study by observations on the structure and lithologic characteristics of the interior of the graywacke districts, as well as of the adjacent formations.

1. Rensselaer Graywacke Upland

General description.—As one approaches the graywacke upland that separates the Hudson valley lowland from the Taconic Range, he travels for the most part over slates in which here and there Lower Cambrian and Ordovician fossils have been found, but none of Middle or Late Cambrian age. Discontinuous, narrow zones of limestone are associated in places with Ordovician slates, and in the Lower Cambrian argillaceous rocks are found occasionally limestone pebble conglomerates, the pebbles of which have furnished most of the fossils of this age. Siltstone zones and thin quartzites are also widespread in the Lower Cambrian rocks. Detailed descriptions of these rocks east of Albany and Troy are by Ruedemann (1930, p. 72–123) and Dale (Dale, 1904a, p. 13–39).

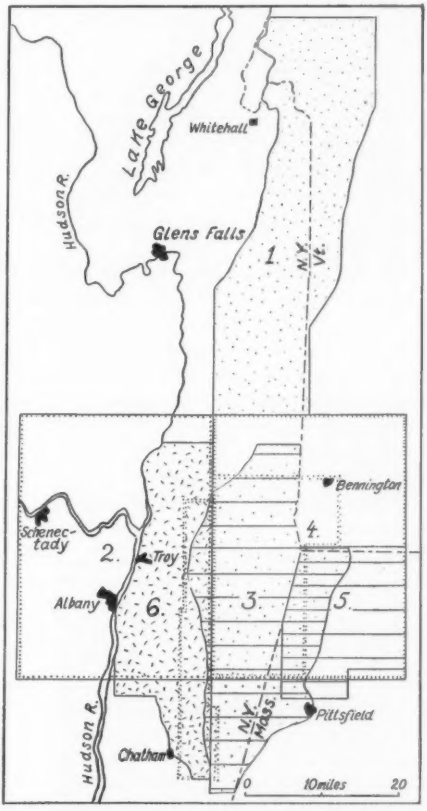

FIGURE 2.—OUTLINE MAP OF TACONIC AREA EAST OF TROY, NEW YORK, SHOWING BOUNDARIES OF AREAS PREVIOUSLY MAPPED

1 (dots): Dale (1899); 2 (blank square with row of dots along border): Ruedemann (1930); 3 (irregular area, dotted, and ruled horizontally): Dale (1904a); 4 (blank square, short dashes along border): Prindle and Knopf (1932); 5 (irregular, ruled area): Pumpelly, Wolff, and Dale (1894); 6 (short dashes): Dale (1893). Area shown in Plate 1 is shown by double row of fine dots.

Rocks of this kind crop out at the slopes leading to the upland, and only at the higher elevations are there layers of graywacke. Here they are interbedded with greenish-gray silty slate, more rarely with purple slate, and to the writer's knowledge nowhere with black slate or limestone. Oddly, however, small masses of limestone crop out in several placed within a few hundred feet of the graywacke. Along the eastern edge of the graywacke upland, lenses of laminated, fine-grained tuffaceous rocks are interbedded with the slates, and the graywacke itself is cut in a few places by basic igneous rocks.

On the upland surface, till is so extensive that only the tops of thicker graywacke zones rise above it. There is reason to believe, however, that slate constitutes much more of the bedrock than does graywacke, and although the word, graywacke upland, is apt, it indicates merely an area where most of the visible outcrops are graywacke, but not that graywacke is the most voluminous bedrock type.

Thus the delimitation of the graywacke area is difficult. We proceed to a description of the border zone, taking up the northern border first, then the western and eastern zones. Important localities are numbered consecutively for easier reference.

Northern border.—The northeastern corner of the upland, near Babcock Lake (Pl. 1), is adjacent to greenish-gray and black slates whose attitudes suggest a broad, southward-plunging syncline. Lenses of limestone, 1–9 feet thick, are interbedded with the slate in a few places (low hill 1160′, 1 mile west-northwest of North Petersburg; local road, 3/10 mile southwest of North Petersburg; west side of State highway No. 22, ½ mile north-northwest of North Petersburg, North Pownal quadrangle, N. Y.-Vt.). Their lithology is said to resemble Deepkill terranes (Rousseau H. Flower, personal communication), but fossils are absent. Two larger lenses of limestone were quarried, years ago, just east of the crest of a low spur, east of Shingle Hollow, 1¼ miles southwest of Hoosick (North Pownal quadrangle, about due east of the letter "w" of the word "Shingle Hollow"). The graywacke appears to overlie the slate and siltstone and is interbedded with it.

Loc. 1. At the summit of *hill 1380′*, 1.1 miles east of Babcock Lake (Grafton quadrangle), beds of medium- to fine-grained graywacke, 1–3 feet thick, are exposed in numerous ledges, dipping 30–40° S. Directly north of the top ledges, interbeds of gray sandy slate between the graywacke layers increase northward until concealed by till, about 120 feet below the summit. The layers appear to be undisturbed.

Loc. 2. *Small rise 1240′*, ¼ mile northeast of the northeast corner of Babcock Lake (Grafton quadrangle). The flat top is underlain by several ledges of white, finely crystalline limestone (Pl. 1, cross section 20). Bedding planes dip at about 20° S., and are strongly lineated in an east-southeasterly direction. Ten feet north of the limestone ledges, a considerable mass of greenish-gray and purple slate displays innumerable small folds and contortions whose axes plunge generally south, or southeast, under the limestone. The contact between the two rocks was exposed in a shallow excavation. It dips at about 20° SW., and the limestone rests on the slate. Although there is no brecciation along the contact, the rocks do not adhere to each other and are fractured. Presumably it is a fault contact. No fossils are known from either the limestone or the slate.

PLATE 2.—VIEWS OF TACONIC AND GRAYWACKE AREAS

FIGURE 1.—VIEW NORTHWARD FROM HILL 1560, 1 MILE NORTHWEST OF THE DOME, BENNINGTON QUADRANGLE, VERMONT

Ridge on right is west front of Green Mountains, culminating in Glastonbury Mountain. Lowland in middle ground is limestone valley of western Vermont. Behind it, in far distance, the Taconic Range of southwestern Vermont, with Mount Equinox. Summit at left is Mount Anthony. Pownal upland makes high ground in left middle distance, and Taconic Range west of North Pownal makes sky line on extreme left. Ledges in foreground are Cambro-Ordovician limestone and dolomite.

FIGURE 2.—LOOKING WEST FROM MANN HILL, BENNINGTON QUADRANGLE, VERMONT

In foreground quartz-chlorite schist of Pownal upland. Dark, forested ridge in center is limestone tract of North Pownal. Hoosic River valley in middle ground, Taconic Range in distance.

FIGURE 3.—VIEW SOUTHWEST FROM MOON HILL, ACROSS PETERSBURG-BERLIN VALLEY

Level sky line, in distance, is Rensselaer graywacke upland, with Sugarloaf Hill in middle distance near center. Slopes on left are western foothills of Taconic Range. Lower Cambrian slates underlie Moon Hill and hill on right edge of view. State Highway 22 and village of Petersburg in valley.

FIGURE 2

FIGURE 4

FIGURE 1

FIGURE 3

GRAYWACKE, SLATE, AND GRAYWACKE DIKE

VIEWS OF TACONIC AND GRAYWACKE AREAS

FIGURE 1

FIGURE 3

FIGURE 2

Graywacke ledges, forming the northern edge of the upland, stand 350 feet south of the limestone, and about 100 feet higher. Interbeds of gray siltstone dip 20°–45° S. Till covers the interval to the limestone knoll, but, a few hundred feet west of the latter, a seemingly continuous section of grayish and purple slate underlies the graywacke.

Till conceals all the important contacts at this locality, but all the rocks have about the same strike and dip. A fault may, or may not, be concealed between the graywacke upland and the limestone. Thus, the slate and the limestone could be Ordovician and underlie the graywacke conformably. Or the slate and the graywacke could be Lower Cambrian, and the limestone Ordovician, originally deposited on slate, and the floor subsequently disturbed by a fault.

For 2½ miles northwest of Babcock Lake, the graywacke layers strike northeast as does the slate in the lowland northeast of the scarp. This involves a peculiar problem. The elevation of the upland is probably caused by the superior resistance of the graywacke over slate. In this section, however, the border of the upland trends northwest, about at right angles to the strike of the graywacke. The lower slopes are till-covered, and there could be a northwest-striking fault that has brought contrasting lithologic phases alongside each other. Before one can make this assumption, however, another matter must be considered. Plate 1 shows that nearly all the dips exceed 45°. Axes of folds are not exposed, and, if the graywacke is simply regarded as a synclinal basin, one obtains a thickness for the graywacke of 3 to 5 miles. Obviously, this is an immense thickness for a formation that covers no more than 24 by 8 miles. If a fault is assumed to cut across this plate on the northeast side, it requires a vertical component of several miles to bring a series of slates to the level at which we now find them. It may not be necessary to assume that the graywacke is abnormally thick. Exposures on the upland are so low and small that axes of folds may very well be numerous but cannot be seen; or the graywacke may be a series of fault blocks, repeated by imbricate thrusts. These assumptions would reduce the thickness of the graywacke and the displacement on the hypothetical fault. However, there is so little indication of a northwest-striking fault that another possibility should be considered. Perhaps the area of graywacke deposition reached its original northeastern limit not far from the present boundary of the graywacke. In that case, most of the graywacke layers would lens out northeast of the scarp and be replaced by the greenish-gray siltstone and slate with the same strike and dip. Before the relative merits of the various explanations can be considered, we may examine additional contact exposures.

Western border.—At several places along the western border, the Rensselaer graywacke is unusually coarse, as is also the Austerlitz graywacke. Where exposures have depth, they indicate that the underlying siltstone-slate formation develops more quartzite lenses along this zone than elsewhere. Practically everywhere the rocks are disturbed, but, apart from the fact that the folds in the thin-bedded siltstone-slate-quartzite series are smaller than those seen in the graywacke, no evidence of a break between the two rock groups was found. The writer therefore assumes that the contrasting dimensions of folds are due to the common correlation between wave length of folds and thickness of competent beds. Exposed faults cut through slate, quartzite, graywacke, and

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PLATE 3.—GRAYWACKE, SLATE, AND GRAYWACKE DIKE

FIGURE 1.—GRAYWACKE LAYERS, 2 TO 6 INCHES THICK, INTERBEDDED WITH GREENISH-GRAY SILTSTONE AND SLATE, DIPPING NORTHEAST

Top of hill 1580′, 1.6 miles north-northeast of School No. 8, North Pownal quadrangle, N.Y.-Vt.

FIGURE 2.—GREENISH-GRAY SILTSTONE AND SLATE, GENTLY FOLDED, WITH FEEBLE AXIAL-PLANE CLEAVAGE DIPPING 70° EAST

On ridge 1260′, south of Sugarloaf Hill, 7½′ Berlin quadrangle, N.Y.-Mass.-Vt. Looking north.

FIGURE 3.—VERTICAL 2-FOOT GRAYWACKE DIKE, CUTTING FLAT-LYING, MASSIVE GRAYWACKE BEDS

On State Highway 2, ¼ mile southwest of Quackenkill settlement, Cohoes quadrangle, New York.

FIGURE 4.—GRAYWACKE, INTERBEDDED WITH SILTSTONE SEAMS, DIPPING WEST

In abandoned road-metal quarry, ¼ mile southwest of East Nassau, Troy quadrangle, New York. Looking southwest.

all other types of rock. Localities are described in order from north to south.

Loc. 3. *Hill 960′*, 2¼ miles north-northeast of the settlement of Quackenkill (east border of Cohoes quadrangle), projects westward beyond the northwest-facing scarp that here marks the northwesternmost section of the graywacke upland (Pl. 1, cross sections 22, 23). Masses of red and purple slate on the north and west slopes are interbedded with lenses of green, gray, and brick-red sandstone, arkose, and, more rarely, of typical greenish graywacke. The entire series, penetrated by quartz veins and narrow breccia zones, is thrown into folds whose axes plunge at about 20° S. and SE. In places, fragments of rigid arkose and graywacke have been thrust across the steeply tilted slates. Obviously, the rocks are very much disturbed, but it is exceedingly difficult to interpret the nature of the disturbing movements. Less than ½ mile northwest is a large ledge of greenish slate in which the layers lie virtually flat, apparently unfolded, with slip cleavage in the regional eastward-dipping orientation. Similarly, at the southeast base of hill 960′, graywacke beds, interbedded with greenish sandy slate, dip at 55° SE., and can be followed for 400 feet up on the scarp. There is a suggestion that the disturbed rocks mark the trace of a high-angle fault of east-west or northeast-southwest strike for the north slope of the hill ends more abruptly than the topographic map suggests, along a straight east-west line that can be followed eastward for ½ mile into the large graywacke scarp, where it appears to die out. Nearly continuous outcrops in the large scarp, east of hill 960′, show graywacke interbedded only with greenish siltstone and slate, whereas, at the northwest base of the scarp, red and purple slate are prominently developed, dipping under the graywacke. Therefore, one might also assume that the scarp is terminated by a northeast-trending fault that has raised the northwestern block, bringing red and purple slates to the level of the lower graywacke beds. In any event, graywacke occurs in both series of slate, sparingly in the red slates, abundantly in the greenish silty sequence.

Loc. 4. *Hill 1260′*, ⅜ mile northwest of Quackenkill settlement (15′ Cohoes quadrangle, N. Y.). West of the summit, a large abandoned quarry exposes a massive layer of partly conglomeratic graywacke, over 75 feet thick, resting on purple slate that contains several lenses of cross-bedded silty sandstone, graywacke lenses, 2–3 feet thick, and greenish silty slate. The indented depositional contacts of all these rocks are repeatedly exposed, slivers and lenses of slate occur in several sandstone beds, and there is no suggestion of a break between the argillaceous and arenaceous rocks. The crest of the hill strikes northeast although, in the vicinity of the quarry, the bedding strikes northwest. As the coarse graywacke underlies the higher portions of this hill, there is a suggestion here that a northeast-striking high-angle fault may be responsible for the abrupt termination of the upland along this hill crest. Exposures on the northwest slope are scarce, but all are of purple slate, and, where bedding can be distinguished from cleavage, the dips are to the east, under the graywacke.

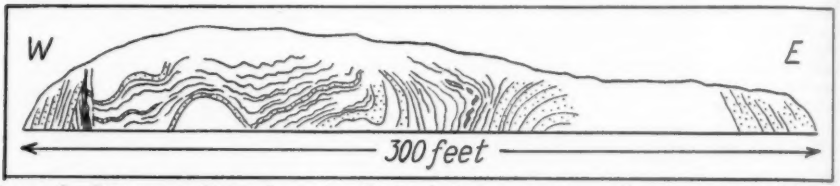


FIGURE 3.—PURPLE AND GREEN SLATE (parallel lines), INTERBEDDED WITH GRAYWACKE (dotted layers) Showing folds and faults. Axes of folds plunge gently south. Cut on State Highway 2, at Quackenkill settlement, Cohoes quadrangle, N. Y.

The thick layer of coarse graywacke that has been quarried at Quackenkill was probably continuous with the similar rocks that underlie hill 1180, ½ mile southeast of the quarry, where the beds dip gently southeast. However, this particular area is evidently much faulted. Several northwest-trending breccia zones, with spectacular veneers of red hematite, and measurable displacements of as much as 9 feet, are exposed in graywacke on both sides of Quacken Kill. The peculiar graywacke dikes, seen in the same zone, are described later. It is impossible to determine details of the structure because the graywacke lacks stratification. Half a mile south of hill 1180, the surrounding purple slate strikes north-south in vertical position, but farther southeast the beds dip gently southeast and are folded and faulted where exposed in highway cuts (Fig. 3).

For about 4 miles south of the quarry, the western border of the upland is fairly straight, and the geological succession is normal. Purple and greenish slate, dipping 20°–40°E., underlies the edge of the graywacke upland on the west. Where bedding can be determined in the massive graywacke by occasional interbeds of slate, a few folds with horizontal axes in a north-south direction can be recognized (Pl. 1, cross section 25). Where road cuts or deeper valleys afford better outcrops, lenses of gray,

white, and greenish quartzite and graywacke are seen within the western slate series. No hiatus in deposition is recognizable between the underlying slates with interbedded quartzites, graywacke, and siltstone layers, on the one hand, and the overlying series of sandy slate and abundant graywacke beds. This was also Dale's conclusion, although he placed the underlying slates in the Lower Cambrian and the main graywacke series in the Silurian (Dale, 1904a, p. 41, 42).

Loc. 5. Dale refers particularly to a *road cut, 2½ miles southeast of Cropseyville* (Troy quadrangle; Pl. 1, cross section 26), where a local road crosses the edge of the graywacke upland, 1 mile northwest of Davitt Pond. The writer's observations agree with Dale's in that there is an apparently uninterrupted succession, from west to east, of reddish and greenish-gray slate with thin, greenish quartzite and graywacke layers, to the main graywacke sequence with numerous beds of purple and greenish silty slate. Thin sections from two arenaceous interbeds in the purple slate series show the typical, slightly recrystallized chlorite matrix between the clastic grains, as it is found throughout the graywacke areas, and both potash feldspar and plagioclase are seen as fragments (Pl. 8, fig. 1).

In discussing the lithology of the terranes west of the Rensselaer graywacke, Dale (1904a, p. 22, 23, 29, 37, 38) speaks repeatedly of "small quartzites". These are lenses of light-gray to white-weathering, fine- to medium-grained rocks, in which quartz grains, ½-2 mm. across, are recognized with the naked eye. At Loc. 5, the quartzites range from 2 inches to over 7 feet thick; on the southeast bank of Kinderhook Creek, ¾ mile northeast of Riders Mills (Kinderhook quadrangle), a lens measures about 20 feet thick; another one (west and northwest of road intersection, 935′ elevation, 1 mile southwest of Red Rock settlement, Kinderhook quadrangle) measures 28 feet, and there are probably even thicker lenses in the terrane. Fresh surfaces are commonly green, or greenish gray, and under the microscope some chlorite is nearly always found. In some, the amount of chlorite exceeds that of sericitic muscovite, and only the amount of quartz in these rocks seems to be more abundant in comparison to feldspar than in typical graywacke specimens. Thus, if the graywacke is to be separated from this series of slates with thin quartzites on the basis of a different composition, the boundary would have to be drawn—somewhat arbitrarily—where the quartz-feldspar ratio passes a significant value. To establish this would require more extensive collecting and mineral counting than was possible within the scope of this study.

Loc. 6. *Gorge of Poesten Kill*, between Barberville and Poestenkill village (15′ Troy quadrangle, N. Y.; cross section 24). This locality was described by Dale (1893, p. 326, 327; 1904a, p. 40) and Ruedemann (1930, p. 150, Figs. 28, 29, 68). Excellent exposures in the stream display interbedded typical graywacke and greenish and purple slate beds, folded with horizontal north-south axes, and cut by high-angle faults striking northwest. However, the westernmost exposures, in the stream bed, are graywacke beds showing coarse quartz-feldspar fragments, and dipping 25° E.; glacial kame gravel and sand conceals the underlying rocks, so that the base of the graywacke series is not exposed. The outcrops show that near the western border graywacke and slate are interbedded through a considerable thickness through no accurate figure can be given as marker beds are lacking, and the structure is too much faulted.

For about 8½ miles south of Poestenkill, graywacke is poorly exposed along the western border of the upland. To the west, largely concealed by Quaternary deposits, are slates with scattered interbeds of quartzitic rocks, folded, and with the limbs dipping generally east. In places, as 1 mile northeast of Sandlake (15′ Troy quadrangle, N. Y.), the westernmost graywacke outcrops show individual, thin layers, interbedded and conformably folded with slate and siltstone beds, or the graywacke faces west as one very thick and relatively coarse layer which commonly underlies the top of a prominent hill with a steep bluff on the west side, as on Bailey Mountain (15′ Troy quadrangle, N. Y.), where a graywacke layer more than 100 feet thick, with pebbles exceeding ½ inch across, rests on greenish slate and gray, sandy siltstone. Pikes Hill, 1 mile south of Bailey Mountain, has a similar position.

Loc. 7. A small outcrop of unfossiliferous limestone has been reported by Dale *south of Bailey Mountain* (Dale, 1893, p. 311 "south of Pikes Hill, near Alps, Nassau [township]"; "Cole farm"). No bedrock was found anywhere near the former Cole farm (now abandoned; the house is shown on the 15′ Troy quadrangle, ¾ mile southeast of Pikes Pond). About 600 feet east of the farm house are ledges of graywacke, interbedded with siltstone and slate. Outcrop locations and observations by Dale are nearly always reliable, but much of the collecting in and around the upland was by J. E. Wolff, whose descriptions are quoted. Perhaps an error has entered in transferring notes. The matter is of some importance for Ruedemann's map of the Capital district (Ruedemann, 1930, map in pocket) shows a long, narrow tract of Ordovician limestone connecting the locality with another one, to be described presently. We have found no outcrops of limestone along this line. In fact, near the middle of the supposed limestone belt is an abandoned quarry (directly east of State highway No. 66, 9/10 mile north of Hoag Corners, Troy quadrangle) showing graywacke and interbedded silty slate. Exposed bedrock on the middle and lower elevations of Pikes Hill is slate, and nowhere have we noticed any limestone debris.

Loc. 8. *The Coonradt farm fossil locality* (little knob, 860′, ½ mile east of Tackawasick Lake; the Coonradt farm is shown on the quadrangle map, just east of State highway No. 66, 15′ Troy quadrangle; Pl. 1, cross section 30). This important out-

crop has often been described (Dale, 1893, p. 311; 1904a, p. 21; Ruedemann, 1930, p. 115–117). A lens of white limestone, with Trenton fossils, measuring 200 by 30 feet, rests on the west on greenish-gray, sandy slate with exposed, interbedded contact. Both rocks dip 60° NE. The knoll is surrounded by till and alluvium, and about 200 feet east-northeast of the limestone are the westernmost outcrops of graywacke, located at the west base of the scarp that leads to the edge of the upland. The rocks here dip 30°–40° E. About 120 feet northwest of the limestone mass a low ridge is underlain by purple and green slate, with a prominent light-gray quartzite, about 30 feet thick, along its eastern border. This rock makes a low wall, dips about 50° E., and terminates the slate complex west of the limestone. Within the slate, the structure is confused. Interbeds of greenish quartzitic graywacke, 1–3 inches thick, dip west and east; in places they can be followed around the apices of gently northward-plunging folds.

At the last-mentioned locality, the western contact of the graywacke upland veers to the southeast and east and is concealed by till. Several smaller graywacke outliers follow to the south, but before they will be discussed, the eastern border and the interior of the upland will be described.

Eastern border.—On the east side, the graywacke adjoins dominantly greenish slate with numerous thin siltstone layers, but with fewer quartzites than on the west side of the upland. Red and purple slate was found at one locality only (east slope of ridge, ½ mile southwest of School No. 3, North Pownal quadrangle). Coarse, conglomeratic graywacke has not been seen on the east side; even medium-grained beds are rare. At several places, graywacke appears to be interbedded with the underlying argillaceous rocks, and outcrops of tuffaceous, laminated rocks just below the edge of the graywacke upland appear to belong to a throughgoing horizon. Disturbances along the eastern border of the upland introduce complications, however; critical outcrops are described in detail, from north to south.

Loc. 9. The northeasternmost outcrop of graywacke is on the west slope of *hill 1460'*, 1½ miles southwest of North Petersburg (North Pownal quadrangle), about 600 feet west-southwest of the single farm at the end of a north-south gravel road. Graywacke layers, 9 feet thick, are interbedded and floored by greenish-gray siltstone, dipping 20° S. One hundred feet east, the highest ledges at the hill expose an interesting series of quartz-sericite-chlorite-epidote-albite schists, about 50 feet thick, well bedded, and criss-crossed by bright-green epidote-quartz veins. The beds dip steeply east and southeast, and axes of small folds plunge about 35° SE. The epidotic schists are underlain by greenish slate, and within the latter are at least three lenses of dolomite, about 15 by 25 feet, folded conformably with the slate. One dolomite mass, about 4 inches across, is completely enclosed by the slate, as seen in cross section. On the north slope of the hill, this group of slate beds rests on southward-dipping crenulated beds of a quartz-sericite-chlorite schist, with innumerable small lenticles of quartz, and stratigraphically below this group of beds, which are continuous with the epidote-albite schist, are ordinary greenish-gray, slightly glossy slates, apparently like those interbedded with the graywacke, but till covers an interval of about 200 feet.

Individual outcrops of the green epidote schist are hard to interpret. However, the association of the rock with highly albitic lenses, the lenticular structure of some zones, the development in places of abundant crystalloblasts of magnetite and carbonate, the association with possible pyroclastics, and the existence elsewhere in the region of albite-rich igneous rocks suggest that these rocks are altered tuffs. Additional occurrences will be mentioned below.

At this locality, nothing definite can be said about the relationship of the graywacke to the epidote schist and the slate that contains the dolomitic lenses.

The structural relations south of Locality 9 call for some comments. The graywacke near the eastern border dips 20°–40° S. and SE. (Pl. 1). Exposures are fairly good, but apices of folds were not seen. As a whole, the graywacke suggests a southeastward-plunging syncline. The structure of the argillaceous rocks east of the graywacke border, on the west slopes of Little Hoosic River valley, does not seem to conform to the synclinal structure of the graywacke. Yet the slates north-northwest of Petersburg show fairly consistently axes of folds plunging southeast. These very incompetent, thin-bedded rocks have been thrown into hundreds of very small folds, and the limbs commonly dip at about 35° NE. The scale of Plate 1 is too small to illustrate satisfactorily the structure of rocks of this type. However, both the graywacke and the slates have been folded about axes in nearly identical orientation. That both rocks are conformably folded is seen at the next locality.

Loc. 10. *Hill 1580'*, 1.6 miles north-northeast of School No. 8, and 1 mile south of Church Hollow (North Pownal quadrangle), is underlain on the east by greenish-gray, siliceous slates, locally with whitish, quartzose lenticles which are albitic in some places and probably of tuffaceous origin. Axes of folds, as well as a strong lineation, just south of the summit, plunge at about 20° SE., and the prevailing dips of the limbs are northeast and east. In the highest outcrops, near the top, there

appear several layers of fine-grained graywacke, interbedded with the slate (Pl. 3, fig. 1). The easternmost ones are a few inches to 2 feet thick, but, within 200 feet southwest, layers as much as 9 feet thick appear in what is evidently a continuous section. The easternmost layers dip east, with the surrounding slate. Farther south, the dip is to the south, apparently caused by one or more folds that can be traced through the outcrops south of the summit. South of these, graywacke and interbedded greenish-gray slate and siltstone beds dip south or south-southwest (Pl. 1). Thin layers of graywacke are found within the slate beds for about ¾ mile south of this hill (north-south trending ridge, ½ mile east of School No. 8), and, although the structural contrast between the vigorously contorted and flexed slates and the much more uniform strike and dip of the graywacke west of it is striking, we believe this also is due to the contrast in the mechanical properties of the two rocks.

Along the graywacke border, between localities 9 and 10, there are at least a dozen small limestone lenses. The smallest one is the one described at locality 9. The largest ones measure about 50 by 20 feet, and up to 9–10 feet thick, constituting east-dipping lenses. Contacts with the surrounding argillaceous rocks are not exposed although four masses are within 2 or 3 feet from chloritic argillaceous beds. One lens, exposed on the northwest slope of Church Hollow (North Pownal quadrangle), ¼ mile northwest of the westernmost house in the hollow, is overlain by greenish-gray siltstone, or graywacke, with quartz fragments as much as 3 mm. in diameter.

Most of the limestone lenses are intensely folded; calcitic, thin-bedded layers especially have been sheared, smeared out, and wrap around the borders of fractured cream-colored dolomitic blocks, as commonly seen in the intensely deformed limestone zones of the Berkshire Hills. As the location and structure of these limestone lenses answer approximately the conditions pictured by Prindle and Knopf (1932, p. 294, Fig. 3, sections B and C)—though these authors do not specifically refer to any of the outcrops here mentioned—the main Taconic thrust may pass directly above them. Nevertheless, the surrounding slate and graywacke are conformably interbedded, and the extremely intense deformation of some of the limestone lenses is perhaps another expression of the unequal behavior of rocks of different mechanical properties exposed to the same stress. Supporting evidence for this possibility is cited in the description of the limestone masses of Butternut Hill, North Pownal, and others.

The contact of the graywacke, south of the last-mentioned localities, is till-covered for about 2 miles, but west and southwest of Petersburg are several important sections.

Loc. 11. *Hill 1454′*, 1.3 miles west-northwest of Petersburg (North Pownal quadrangle). State highway No. 2 passes the eastern border of the graywacke area at the northern base of this hill, and, at an elevation of 1140 feet, several cuts expose horizontal graywacke layers, conformable on dark-gray slate, and the interbedding of the two rocks is well displayed. No graywacke has been found at lower elevations on this hill slope, and, at the summit, flat-lying graywacke is interbedded with gray, silty slate. At the southeastern base of this ledge is a group of well-laminated, fine-grained siliceous greenish beds, some of which exhibit abundant sparkling rhombs of calcite, and magnetite octahedra, 1–4 mm. across (Pl. 8, fig. 2). Thin white veins and irregular pods are composed of quartz and albite, and this group of rocks is considered another tuff lens, interbedded with the silty slates that underlie the graywacke. The exact composition of the albite is difficult to determine. The grains are rarely larger than .2 mm. and so intimately mixed with quartz that the indices of refraction are hard to determine, especially as some of the ubiquitous chlorite and sericite of the rock forms minute specks and dust films around most of the grains. In all the thin sections examined, the indices of the plagioclase are lower than balsam, and in twinned crystals extinction angles up to 19° were measured. Therefore, the composition is assumed to be that of sodic albite, perhaps pure albite in places. The albite-bearing rocks range through some 30 feet of the section, and their boundaries, as far as the imperfect exposures permit one to judge, are gradational with the surrounding slates. A lens of ash, deposited with silty and muddy sediment, might produce these relations, and Tilley's term, "schists of composite origin" (Tilley, 1923, p. 188), seems applicable to these rocks. It is interesting that this lens, like the one at localities 8, 11, and 13, occupies a position slightly below the apparent base of the graywacke formation.

Loc. 12. *Sugarloaf Hill* (7½′ Berlin quadrangle, N. Y.-Mass.-Vt.), and the *ridge 1300′–1260′*, ¼ mile southwest of it (eastern border of 7½′ Taborton quadrangle, N. Y.), display one of the best sections on the eastern border of the graywacke upland (Pl. 2, fig. 3). Here, through a vertical relief of about 600 feet, is exposed a great mass of greenish-gray arenaceous and silty slates, with many interbeds of gray, argillaceous sandstone and, near the summit of Sugarloaf Hill, a few lenses of conglomerate in which fragments of light-gray or greenish, fine-grained quartzitic rocks range up to 5 inches across. Although open folds and eastward-dipping cleavage are seen in the thinner argillaceous beds (Pl. 3, fig. 2), the thicker siltstone and sandstone layers are about horizontal or show merely gentle undulations. These instructive outcrops suggest that the degree of deformation of the rocks in this section depends to a large measure on the competency and thickness of each group of layers. Isoclinal folds were looked

for in vain, and there are extensive ledges where the hill surface coincides with the gently flexed, but nearly horizontal beds of stronger sandstones and siltstones, like small plateaus. The tops of these two hills are about 200 feet below the level at which the graywacke upland breaks off, and the interval, about 900 feet horizontally, is poorly exposed though a few low outcrops exhibit the same sort of greenish slate as is seen in the main slopes of the hills. As one approaches the edge of the upland closely, the graywacke is found resting upon greenish-gray quartzose slates interbedded with a lens of what appears to be a tuffaceous rock, with conspicuous lentils of quartz that could be flattened lapilli. These highest argillaceous rocks dip about 15° W. under the graywacke. Where the bulk of the graywacke appears, till covers the bedrock, so that one can judge the relations only from scattered low outcrops, or the composition of the earth under uprooted trees.

The section just described should pass through the Taconic thrust, and it could be assumed that the fault plane passes the hillside where the till cover conceals the bedrock. Nevertheless, the lithology of the section as exposed on Sugarloaf Hill is remarkably similar to that of locality 10 where graywacke is visibly interbedded with the greenish, argillaceous rocks. Even though, in the writer's opinion, there are no typical graywacke beds among the rocks at Sugarloaf Hill and vicinity, the coarser layers of impure sandstone resemble it closely, and a special examination of the extensive exposures on these two hills may well disclose graywacke that has been overlooked. Supporting evidence that the graywacke is interbedded with the argillaceous suite of rocks east of it, is found at the next locality.

Loc. 13. *State road No. 154* exposes a good section through the eastern border of the graywacke upland, 1 mile west of Berlin, New York. From the top of the upland, at an elevation of 1650 feet, down to an elevation of about 1280 feet, road cuts expose a nearly flat-lying sequence of greenish-gray and black, siliceous slates. Fine-grained graywacke, in beds and lenses, a few inches to over a foot thick, is interbedded with these rocks down to the lowest levels. However, the exposures are not continuous, and, as the road runs about parallel to the strike of the rocks, they do not afford a complete vertical section. In nearly every other outcrop there are slickensides and fault planes, most of which dip 25°–60° E. Striations on them trend east-west, and from the one-sided roughness of the surfaces, and a few displaced quartz veins, it seems that some are upthrusts, others normal faults. Perhaps these are the traces of movements subsidiary to the Taconic thrust. If they are, it is difficult to place the main thrust because graywacke appears throughout this section. The argillaceous beds are thrown into small, westward-overturned open folds. Axial-plane slip cleavage is not present everywhere, but, where seen, the planes dip about 75° E., and are evidently older than the fault plane.

Till covers the eastern border, south of Loc. 13. The obscure relation of the graywacke to the limestone mass south of Berlin is discussed later. Though the argillaceous rocks east of the graywacke are well displayed in the slopes leading down to Little Hoosic River, no sections were seen in which slate and graywacke are interbedded or in fault contact. Commonly a till strip separates the easternmost graywacke outcrop from the westernmost slate. Cross-sections 11–13 illustrate the general structure in this interval. A small lens of white limestone within the greenish slate is exposed near the summit of hill 2070, $\frac{4}{10}$ mile southwest of Rocky Hill (Taborton quadrangle; cross section 12), and, near the border of the upland, low dips are the rule.

The next locality shows the relation between graywacke, slate, and a limestone lens particularly well.

Loc. 14. *Buck Rock* (southeast corner of Taborton quadrangle; cross section 14), $1\frac{3}{4}$ miles west of Cherryplain (Pl. 1), at the eastern edge of the upland, has extensive ledges on both the northeast and east side. On the northeast slope, one can walk from the edge of the upland, underlain by gently undulating graywacke beds, about 100 feet down, and observe the lowermost graywacke beds resting on greenish siliceous, fine-grained slates. Less than another 100 feet east, the slate dips more steeply and displays numerous axes of folds plunging gently north and south. There are light-gray, nearly aphanitic lenses with abundant white quartz-albite veins; microscopic study suggests that the rocks are tuffaceous like the similar rocks described above.

East of the summit, siliceous slates form an east-facing ledge, about 70 feet high, in which lenses of typical graywacke, with quartz and feldspar fragments up to 8 mm. in diameter, are interbedded with the slate. All the rocks are folded. Fold axes plunge about 20° N.-NE., and the dip of the limbs varies so much that one cannot say whether the over-all dip is to the east or to the west. Axial-plane slip cleavage in the more argillaceous zones dips about 80° E. The lowest ledges show low westerly dips, and, at the base of the ledge, the argillaceous beds rest on limestone, with fully exposed gradational contact through about an inch of impure calcareous slate. Bedding is well preserved in both rocks, both megascopically and microscopically, through the interbedded zone. About 4 feet above the top of the limestone are graywacke layers with feldspar grains, 2 mm. across, and 20 feet above are quartz and feldspar fragments over 5 mm. in diameter. About 4 feet of limestone is exposed. The rock seems to form the crest of a small anticline because, about 45 feet east of the contact, slate dips about 60° E.

The contact of the limestone with the westerly

slate-graywacke series can be followed southward for about 600 feet. The gentle northerly plunge of the fold axes makes it rise slightly southward at the cliff front. In one place, about 300 feet from the first-mentioned ledge, a vertical fissure striking east has caused a small re-entrant in the cliff front and permits one to follow the undulating contact of the limestone with the overlying slate for about 7 feet (Fig. 4). South of this, limestone ledges along the east base of the hill are about 120 feet high, but the internal structure of the rocks is obscure. The limestone could be an unbroken sequence of layers, but it could equally well be a thinner section distorted by folds. Some 1000 feet south of the last contact exposure, till covers the ground, and the southern termination of this limestone is concealed.

This limestone cannot be assigned precisely to any particular formation or age. At Butternut Hill, 1½ miles east, is a large mass of similar limestone, very probably Ordovician. At Buck Rock, the bulk of the limestone is not in visible contact with the greenish slate-graywacke series, and no fossils or black slate have been found that might suggest Ordovician age. Like the exposures at localities 2, 8, and 9, this one leaves the age of the graywacke series uncertain.

South of Buck Rock begins a large till-covered area, and no other contacts of the graywacke upland have been found. Black slate lies east of the upland, and, southwest of Stephentown Center, gray, red, and greenish slates are poorly exposed but do not approach the graywacke closely. An outcrop of limestone, said to occur near Taplin Pond, has not been located.

Interior of the graywacke upland.—The graywacke upland is a rolling plain, mostly covered by dense underbrush and second-growth forest; large parts are swampy flats most difficult to traverse except in the winter season. Some of them have been converted into lakes, and a number of camps and tourist colonies have been established, but during the winter the upland is nearly uninhabited.

Through the extensive cover of glacial till and kame gravel, a number of low ridges bring bedrock to the surface. G. K. Gilbert (Dale, 1904a, p. 48) suggested that a mass of Precambrian rocks in the interior may have supplied the coarse debris of the graywacke. Though the upland has been traversed extensively, no Precambrian outcrop was found. Instead, the ridges are underlain by graywacke, and ordinarily the rocks are so massive that their structure remains unknown. In some places, however, slate appears, usually along the west base of the graywacke cliffs; by means of these interbedded argillaceous zones, the

FIGURE 4.—SILICEOUS SLATE, WITH GRAYWACKE INTERBEDS, AT THE EASTERN BORDER OF THE RENSSELAER GRAYWACKE UPLAND

Resting conformably on a lens of limestone. East base of Buck Rock, about 400 feet south of summit. Taborton quadrangle, N. Y.

structure of at least a few graywacke belts appears clear. They are plates, rarely more than 100 feet thick, commonly much thinner, and dip eastward or southeastward. The extent of folds is unknown. Along the borders, where exposures are deeper, the writer has mapped a few. But it is unknown whether the tilted graywacke plates in the interior are the limbs of westward-overturned folds or fault blocks. In the northwestern area, some of the wider slate belts could be mapped, but in the rest of the upland the exposures are too small and scattered. Along the eastern border, some dips are gently west, and to that extent Dale's conception of a synclinal nature of the graywacke (Dale, 1893, p. 324) is correct, but the true structure of the interior remains obscure.

TABLE 1.—CHEMICAL ANALYSES OF GRAYWACKE, SCHIST, AND ALBITE BASALT

	1	2	3	4	5	6	7
SiO_2	71.36	72.62	55.57	69.11	47.37	47.40	49.84
TiO_2	.69	.78	1.00	.60	1.05	2.72	2.79
Al_2O_3	10.48	10.15	23.78	11.38	23.76	10.34	11.57
Fe_2O_3	1.08	0.66	1.50	1.41	2.32	4.44	1.36
FeO	3.81	3.52	6.73	4.64	7.48	11.56	12.89
MnO	.11	.04	.21	.17	.06	.24	.21
MgO	1.61	1.94	1.73	2.06	1.99	5.31	6.71
CaO	1.67	1.34	.24	1.15	.73	10.26	7.56
Na_2O	2.33	1.84	.81	3.20	1.89	2.84	2.90
K_2O	2.31	1.80	3.15	1.76	7.81	1.16	.35
P_2O_5	.05	.03	.05	.03	.09	.24	.30
H_2O^+	3.90	3.75	5.14	4.13	4.69	2.98	3.70
H_2O^-	.04	.04	.04	.05	.01	.19	.10
CO_2	.30	1.05	.00	.00	.55	.00	.00
C	—	.12	—	—	—	—	—
S						.14	.05
NiO						.00	.00
Cr_2O_3						.00	.02
$-S = O$						−.07	−.02
	99.74	99.68	99.95	99.69	99.80	99.75	100.33

1. Coarse graywacke, ¼ mile west of Kendall Pond, 7½′ Taborton quadrangle, N. Y.

2. Graywacke lens in chloritoid-quartz-chlorite schist, southwest spur of hill 2135′, 1½ mile east-southeast of Center Berlin, 7½′ Berlin quadrangle, N. Y.-Mass.-Vt.

3. Chloritoid-quartz-chlorite schist, cut on State Route 346, 1½ miles northwest of North Pownal, Vermont, 7½′ North Pownal quadrangle, N. Y.-Vt.

4. Graywacke, cut in State Route 203, 1½ miles east of Spencertown, New York, 15′ Kinderhook quadrangle, N. Y.

5. Albite porphyroblast quartz-muscovite schist, Hoosac schist. Dump at Central Shaft of Hoosac Tunnel, 7½′ North Adams quadrangle, Mass.

6. Albite basalt, southeast slope of hill 1653′, ½ mile northeast of airway beacon, Turner Mountain, 7½′ Stephentown Center quadrangle, N. Y.

7. Albite basalt, loc. same as No. 6, close to contact with graywacke.

Analyst: H. B. Wiik, Helsinki

Petrography of the graywacke: The lithology of the graywacke has been described in considerable detail by Dale, and the following comments are brief.

Typical Rensselaer graywacke is a dull-green, massive rock, weathering light gray, even white. Quartz grains, clear, glassy, and commonly with greasy luster, and fewer of porcellain-white feldspar are recognized in most outcrops. Joint veneers of quartz, deep-green chlorite, and cream-colored carbonate add color to many intensely fractured ledges of the rock, and surfaces of coarse graywacke, displaying hundreds of red, pink, or white feldspar fragments in the green matrix, are beautiful, and well illustrated in a color plate by Dale (1893, Pl. C). In places, concretions, presumably of manganiferous carbonates, deeply decomposed to powdery masses of black and brown oxides, were noted. Their contacts are gradational, and traces of carbonates are seen in the groundmass of many thin sections of graywacke.

Examination of 28 thin sections, combined with studies of specimens of coarse varieties, gives the following composition of clastic fragments, in order of decreasing abundance: quartz, sodic plagioclase (ab_{80-98}), potash feldspar (microcline fairly abundant, amount of orthoclase less certain), tourmaline, zircon, sphene, epidote, hornblende, pyroxene, apatite, garnet, rutile, magnetite, pyrite, graphite, and ?staurolite. In this list should be included muscovite and biotite, but it is difficult to estimate their amounts because of the involved age relations among the micas. The groundmass is largely an aggregate of colorless mica and light-green chlorite plates which range from minute specks to plates nearly .5 mm. long. Probably, the larger muscovite plates are clastic constituents of the graywacke, and the larger chlorites seem to be clastic biotite flakes in an advanced state of chloritization. Pleochroism ranges

from dull green to pale brown, and the birefringence is commonly a high first order, quite unlike that of true chlorite. On the other hand, there is ample evidence that the groundmass of the graywacke has recrystallized, and larger scales of muscovite as well as stubby plates of chlorite, optically homogeneous,

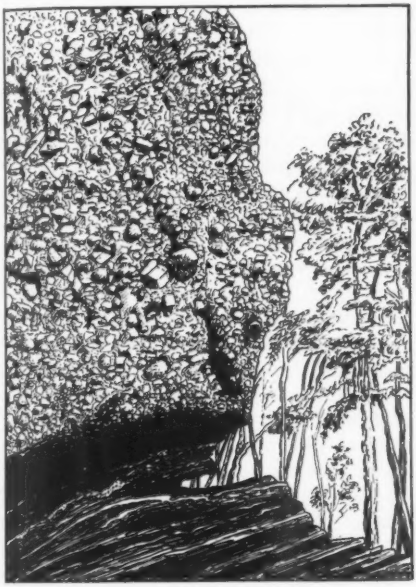


FIGURE 5.—LENS OF CONGLOMERATE, 25 FEET THICK, RESTING ON BLACK SLATE

On western border of Austerlitz graywacke area, 1½ miles east-northeast of Spencertown. Looking south. Drawn after photograph.

and with normal, low birefringence, are believed to be secondary, recrystallized grains. Rhombs of carbonate also replace the groundmass of the rock in a few thin sections.

Three chemical analyses (Table 1) of coarse, medium-grained, and sheared graywacke, the first from the Rensselaer upland, the second from the Austerlitz mass, and the third from a lens about 3 miles east of the east border of the upland, show silica a little higher than Tyrrell's and Pettijohn's averages (Pettijohn, 1949, p. 250) and soda about equal to potash, instead of greater, but the sum of combined iron and magnesia, 6.50, 6.12, and 8.11, respectively, is well within the range characteristic of graywackes. The amount of CaO is appreciably less than graywacke averages, and apparently this graywacke has been derived, at least in part, from rather siliceous rocks. Chemically, it stands close to Pettijohn's "subgraywacke" (Pettijohn, 1949, p. 255).

Two lenses of coarse conglomerate in the graywacke of the Austerlitz area give some information on the source rocks of this mass. One, on the north-northeast-trending crest of double hill 1240′, ¾ mile southeast of Red Rock settlement (Kinderhook quadrangle, N. Y.), is about 12 feet thick but widens near the hill top. Cobbles over 1 cubic foot in size are occasionally seen, and pebbles, 4–5

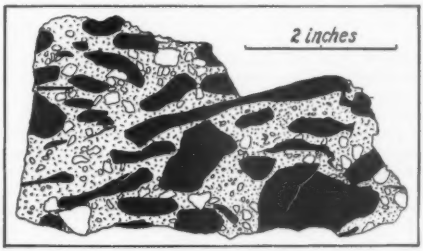


FIGURE 6.—CONGLOMERATIC GRAYWACKE WITH JET BLACK FRAGMENTS

Aphanitic cherty slate, perhaps of tuffaceous origin. One and a half mile northeast of Quackenkill settlement, Cohoes quadrangle, N. Y.

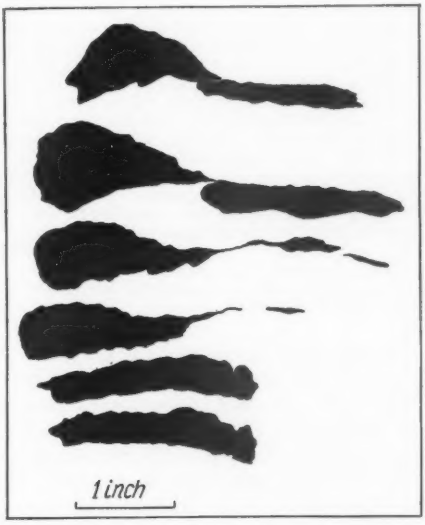


FIGURE 7.—SIX PARALLEL SECTIONS ACROSS BLACK PEBBLES

In specimen of conglomeratic graywacke, showing thin, sheetlike extensions of pebble. Same locality as Figure 6. Sections cover 3½ inches of rock.

inches across, are common. Most of them are vein quartz, a few are gneissic granite and pegmatite, and some are a greenish-white, finely crystalline limestone, with small amounts of clastic quartz, and chlorite. One large block of a hydrothermally altered basalt displays well-preserved diabasic texture (Pl. 8, fig. 3). The second lens is about half way up on the west slope of hill 1360′, 1.5 miles east-northeast of Spencertown (Kinderhook quadrangle).

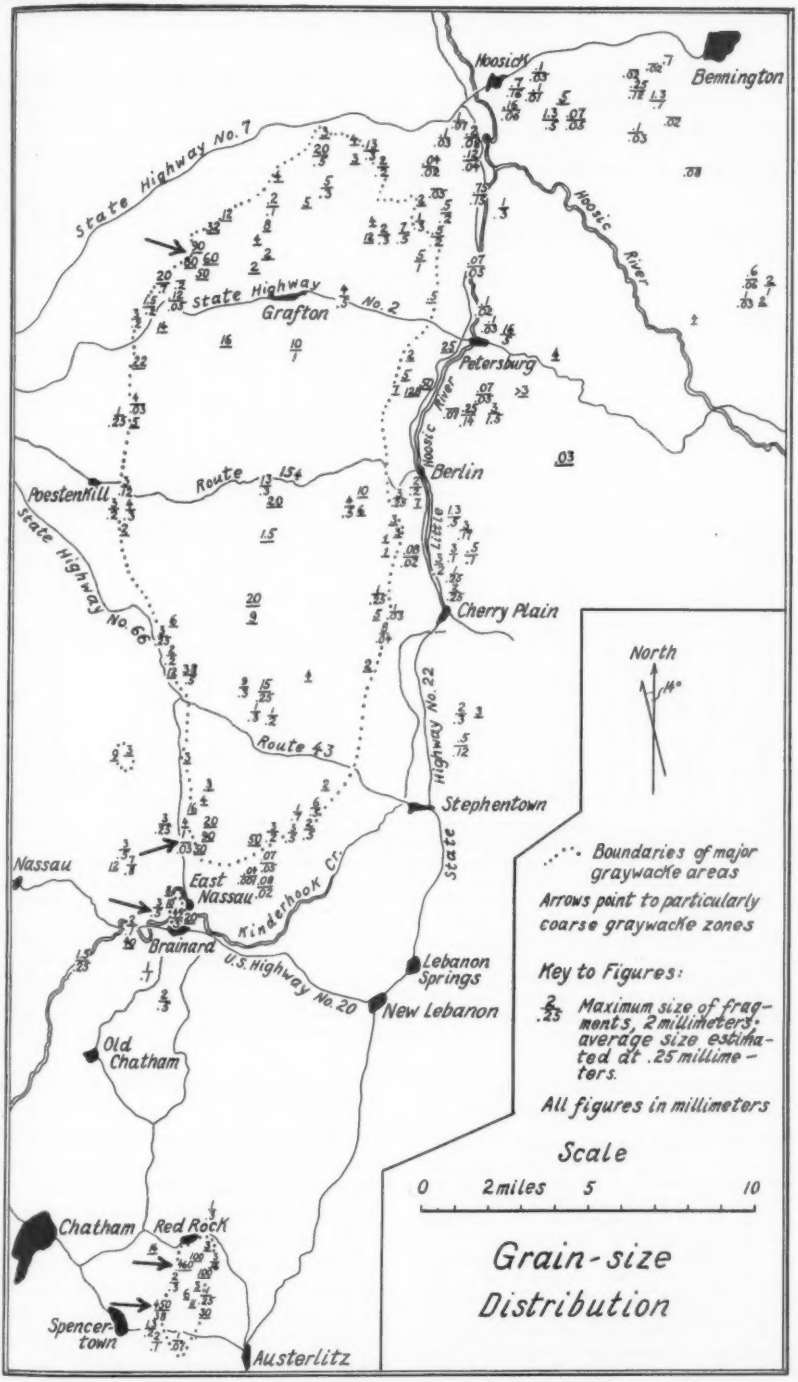
A west-facing cliff exposes in cross section a lens, about 25 feet thick, resting on black slate, and extending along the strike for about 150 feet (Fig. 5). Almost all the cobbles, about 6–8 inches across, are a white, medium-grained, somewhat gneissic albite-muscovite granite (albite ab_{97}), with subordinate chlorite and biotite. Pegmatite is also common, and —apart from the ubiquitous quartz—there are several types of calcareous phyllite and schist. This conglomerate has been deformed plastically, and, while the granitic pebbles have maintained blunt, ellipsoidal shapes, those of phyllites and schists have been distorted into curved sheets and discs, molded around the surfaces of the granitic cobbles which plunge about 30° NE. The base of both conglomerate lenses is conformable with slate, but sharp. The top contacts are not exposed, and along the strike the coarse lenses lose their identity in graywacke of average grain size.

Conglomeratic lenses, about 3 feet thick and 6–10 feet long, in cross section, were noted east and northeast of Hoag Corners, and near the summit of Bailey Mountain (Troy quadrangle). About a mile northeast of Quackenkill settlement (Cohoes quadrangle) the graywacke displays clusters, or swarms, of jet-black, aphanitic pebbles, composed of minute angular fragments of quartz, plagioclase, and ?pyroxene, $\frac{1}{15}$ to $\frac{1}{100}$ mm. in size, embedded in a brownish amorphous groundmass, interspersed with numerous thin muscovite scales, parallel to well-developed bedding (Fig. 6). Thin lenses and extensive films of graphite, parallel to it, pervade most of the pebbles, and some are coated with graphite, commonly showing slickensides and high polish. A specimen with black pebbles was slabbed, and, as the cross sections show (Fig. 7), some pebbles have thin sheetlike extensions suggesting they were embedded in the graywacke before they consolidated. This is also suggested by the irregular contacts shown by the pebbles under the microscope. Quartz and feldspar fragments of the graywacke impinge on the fine-grained rock as if the material had yielded to the larger clastic debris.

Distribution of grain size: The grain size of the graywacke was recorded as often as possible in the field, and data of 189 localities are plotted in Plate 4 and Figure 8. The figures are of limited value because it is difficult to compare a pair of maximum and average figures, the one pair referring to one hand specimen or thin section, the other to measurements on several dozen square yards. Nevertheless, a few figures, even if only of qualitative value, are probably better than nothing. All the coarse phases of the graywacke, as well as conglomerate lenses, are crowded along the western border of both the Rensselaer and Austerlitz areas; it is probably no accident that the thickest section of graywacke and siltstone is at the northwest corner of the upland, 2 miles south of Pittstown (Grafton quadrangle). In the map are also included data for graywacke lenses east of the Rensselaer upland, which are fine-grained except for two localities at, and north of, Sugarloaf Hill (Locality 12), where small conglomeratic lenses display fragments up to 50 and 128 mm. in diameters respectively. These two masses are probably pyroclastics.

GRAYWACKE DIKES: A cut on State Highway 2, opposite a little reservoir, ½ mile southwest of Quackenkill (Cohoes quadrangle), exposes a vertical dike, 2 feet thick, of a fine-grained to aphanitic, greenish-gray rock cutting a flat-lying layer of massive graywacke (Pl. 3, fig. 3). In the field, the rock was thought to be an igneous dike though no contact effects could be seen. Microscopic study shows that the dike is composed of very fine-grained constituents of the graywacke, but sericitic muscovite is much more abundant than chlorite. The principal clastic fragments are quartz, albite (ab_{95-99}), microcline, orthoclase, zircon, sphene, and a little tourmaline. The contacts are obscured by finely divided graphitic dust as well as hematite which pervades considerable portions of the dike and surrounding graywacke. One other, thinner dike of a similar rock was seen in the hillside southeast of this dike, and fractures, breccias, and small northwest-striking faults penetrate this entire section of the graywacke and interbedded slate (Fig. 9).

In the abandoned road-metal quarry, northwest of Quackenkill, many aphanitic, olive-green veins, rarely over ½ inch thick, penetrate the graywacke irregularly. Several displace the adjoining rock, and some are frayed at the ends as if two blocks of graywacke were squeezed past one another. A few branch into numerous fine endings and thin films that merge with the surrounding graywacke; still others connect with quartz-chlorite veins. It is difficult to say whether the graywacke had completely solidified before these fissures formed. They are filled, like the large dike, with exceedingly fine detritus, apparently of the same composition as the graywacke. Quartz fragments of the surrounding graywacke that lay at the vein contacts have commonly recrystallized, projecting as beautiful prisms and rhombohedra into the veins, and incorporating dark, dusty material after the fashion of phantom quartz. Internal contortion of the vein material, lamination, or



MAP SHOWING DISTRIBUTION OF GRAIN SIZE IN GRAYWACKE AREAS, AND OF CLASTIC FRAGMENTS IN LENSES INTERBEDDED WITH SURROUNDING ARGILLACEOUS ROCKS

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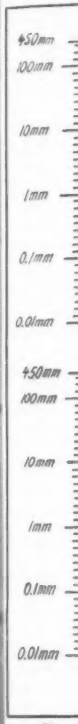

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alignment of debris which Shrock (1948, p. 216) considers suggestive evidence of intrusion from the side or below are generally lacking in these veins. In the writer's opinion, the highly irregular contacts and shapes of the veins, and the close similarity between vein fill and the finer constituents of the graywacke, suggest that at least some of these veins opened and were filled before the graywacke had consolidated.

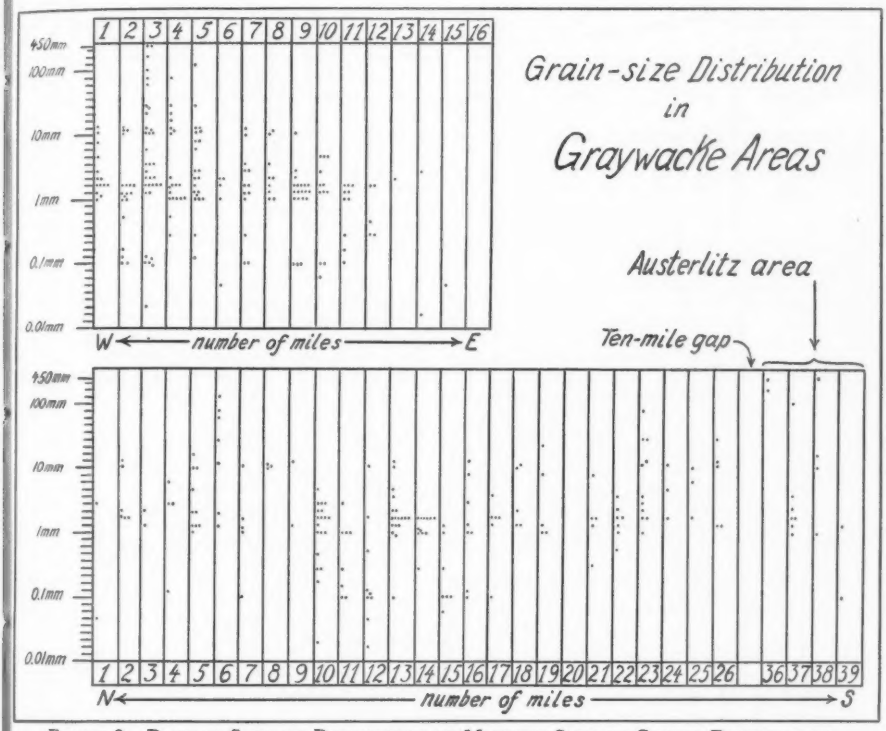

FIGURE 8.—DIAGRAM SHOWING DISTRIBUTION OF MAXIMUM SIZES OF CLASTIC FRAGMENTS IN GRAYWACKE AREAS

Data in upper diagram are arranged in columns showing distribution from west to east. Zero meridian is 3 miles west of eastern border of 15′ Troy quadrangle, New York. Each column contains data of localities lying as many miles east of zero meridian as figure at top of column indicates. Lower diagram shows data of the same localities arranged from north to south. Zero parallel is 1 mile south of northern border of 7½′ Grafton quadrangle, N. Y., and localities are as many miles south of that parallel as figure at bottom of column indicates. No data were collected for 10-mile belt north of Austerlitz graywacke area.

IGNEOUS ROCKS: Brief mention of igneous rocks in the Rensselaer graywacke is made by Wolff and Dale (Dale, 1893, p. 310, 327) and Prindle and Knopf (1932, p. 282, 283), but the localities are obscurely given, and the rocks not described. The writer knows of two areas, one exposing a thin, highly altered basaltic flow, with pillows, and variolitic structure, the other a group of three dikes of albite basalt. We have been unable to locate a few other outcrops mentioned by Prindle and Knopf (1932, p. 282).

Loc. 15 West-northwest slope of *hill 1280′*, ½ mile west-southwest of Banker Pond (Grafton quadrangle). In the till-covered slope there is a single ledge, at an approximate elevation of 1100 feet, that extends horizontally for about 300 feet. This is believed to be what Wolff and Dale termed a "metamorphic eruptive" (Dale, 1893, p. 310, 311) 327). Graywacke of normal composition is here underlain by a lens of an aphanitic, dull-green rock, about 60 feet thick. Eastward, the lens thins until it cannot be distinguished from the graywacke; the western continuation is covered by till. The westernmost exposures are criss-crossed by

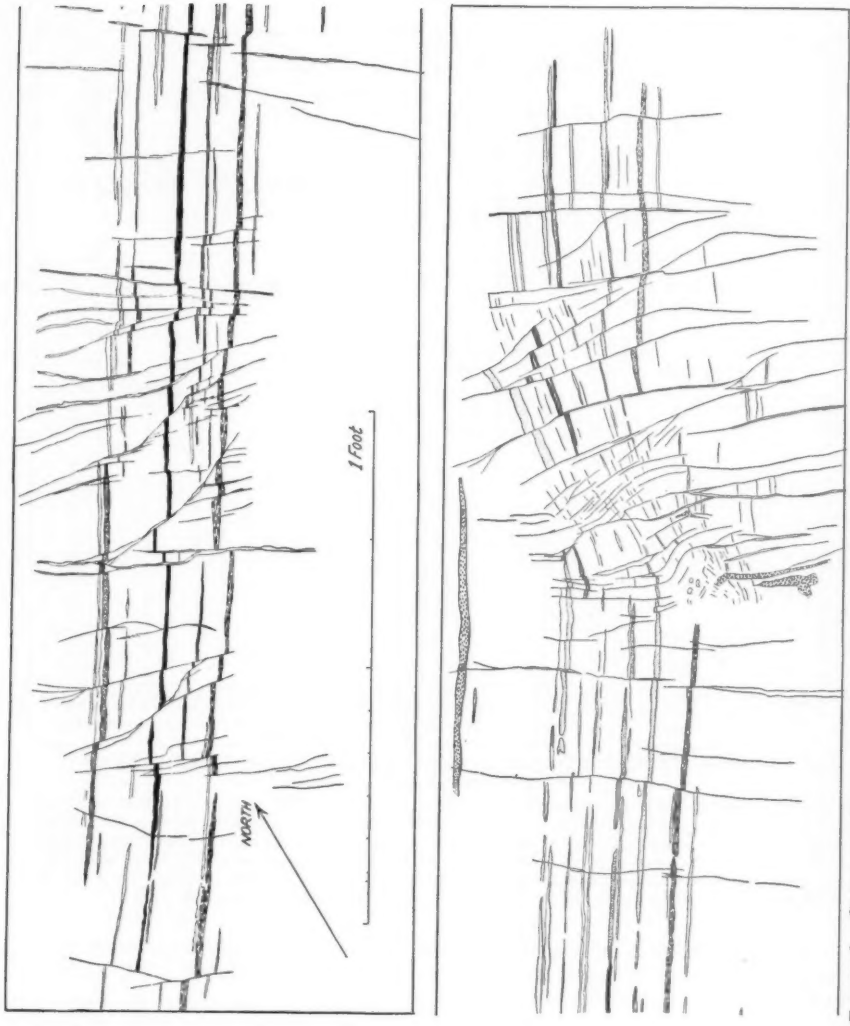

FIGURE 9.—GLACIATED SURFACE OF RED SLATE, WITH THIN INTERBEDS OF GREENISH SILTSTONE AND SANDSTONE (dashes, dots, and circles), SHOWING INTENSE FRACTURING, AND SMALL DISPLACEMENTS

Above, northwest strike; south and east dips. Pasture, ½ mile northeast of Quackenkill set-

bright-green epidote-quartz veins, and this rock is floored by dark-gray and purple slate, dipping 58° SE. In it, too, are several thin conformable lenses of epidote.

At the west end of the ledges, the compact rock contains ellipsoidal pillows, 4–6 inches across; the weathered surfaces of some of them exhibit spherules, or varioles, about 4 mm. across, with a dark-green core and gray shell. Under the microscope, the varioles show bunches of radiating needles, wisps, and laths of a colorless, weakly birefringent mineral, so thoroughly permeated by dark dust particles that identification proved impossible (Pl. 8, fig. 4). On the edges of a few sections, where the dust is not too thick, indices are lower than balsam. Most of the fibers have parallel extinction, though angles of 12° and 16° were measured in a few laths. Scattered through and between the spherules are veinlets of quartz and albite (ab_{98}); these are in turn crossed by innumerable crooked epidote veins. Still later are quartz veins with a little calcite and albite, and the latest veins are nearly pure chlorite. In the cores of almost every variole is also a speck of chlorite with deep berlin-blue interference color. Although polysynthetic twinning is not apparent in the radiating crystals, they are probably a sodic plagioclase although they may be zeolites. The difficulties inherent in the examination of such rocks are well stated by Mügge (1926, p. 50–51), and, in view of the smallness of the spherules, no chemical analysis has been attempted. Some of the radiating groups of crystals appear to have branched out from minerals replaced by the fine-grained secondary material. No pyroxene, calcic plagioclase, or other basaltic minerals have been found, but the pillows, the present mineral composition, and the manner in which the spherules are concentrated in this small lens, suggest that it is a remnant of a much-altered, small, basaltic or andesitic flow that congealed under water.

Loc. 16. *Turner Mountain*, a northeast-trending ridge that terminates the graywacke upland on the southeast, about 2–3 miles southwest of Stephentown (Stephentown Center quadrangle), slopes gently toward the graywacke upland, but, on the southeast, the ridge breaks off with fairly steep slopes. At three places (about 600 feet south of the Airway beacon at Turner Mountain; high on the south slope of hill 1570′, ½ mile east of the beacon; and high on the southeast slope of hill 1653′, 1 mile northeast of the beacon), dikes of a basaltic rock, 30 to 150 feet thick, penetrate the graywacke underlying the ridge and dip steeply northwest, nearly parallel to the graywacke bedding.

Megascopically, the basalt is light greenish gray, fine- to medium-grained, showing here and there matted laths of plagioclase, about 1 mm. across, and a few specks of pyrite. Quartz veins with or without epidote, calcite, pyrite, and chlorite penetrate several ledges which rise as relatively resistant rocks 10 to 15 feet above the level of the surrounding rocks on the hillside. Table 1 (Nos. 6 and 7) gives two chemical analyses of the least altered phases of the rock.

The microscope shows an aggregate of extensively altered plagioclase and pyroxene with diabasic texture in places (Pl. 8, fig. 5), cut by innumerable veinlets of quartz and epidote. Nearly all the pyroxene of some slides has been altered to chlorite and secondary magnetite; in others, common hornblende, chloritized outwardly, fringes the pyroxene grains. The freshest pyroxenes give optical data of common clinopyroxene—$2V(+) = 25°$, 35–45°, 60°; $Z \wedge c = 40°$–48°; $r > v$,— but indices of refraction could not be determined accurately as the minute micaceous dust particles line practically every smallest surface of the grains. In 10 powder preparations and thin sections the plagioclase was determined as albite (ab_{95-98}), and no trace of calcic plagioclase has been found. Thus, the rock might be called a spilite. However, the minerals of the rock are so altered that it is impossible to decide whether the albite was derived from an earlier calcic one or, as assumed by Sundius (1930, p. 1), was the primary plagioclase. Fairbairn's (1934) observations, too, leave the question open.

Like pyroxene, albite in the rock has commonly been altered to a fine-grained felt of sericite, chlorite, quartz, and traces of calcite and magnetite. Ilmenite is generously scattered through the altered portions, as skeletal growths of trigonal symmetry, surrounded by sphene. Leucoxene, epidote, clinozoisite, and specks of hematite are common in the altered phases of the basalt.

The graywacke has been silicified for several inches near basalt contacts. Chlorite of the groundmass has been recrystallized to needles and bunches of a faintly pleochroic actinolitic hornblende which pierce all minerals, including quartz. Clastic feldspar, for the most part a sodic plagioclase, is quite fresh. It may be concluded, therefore, that the extensive sericitization of the plagioclase in the basalt was due to late magmatic alteration which remained confined to the igneous rock.

QUARTZ VEINS WITH ALBITE OR EPIDOTE: Quartz veins penetrate the graywacke by the thousands. Most of them are thin and discontinuous, and their mineralogy simple. A few grains of pyrite, chlorite, and, locally, carbonate are the most common associates. However, closer inspection discloses some grains of albite (ab_{94-100}), 1–3 mm. in diameter, weathering as chalky-white spots amid the more resistant quartz aggregate. The composition was determined at 20 places scattered through the entire area, and there seems to be no regional variation in composition. Albite rarely amounts to more than 1 per cent in the vein rocks, commonly much less. Veins in the southeastern part of the Austerlitz outlier contain pink albite (β = 1.529), simulating potash feldspar. Where it is plentiful, as in a cut on State Route 203, 1½ miles east of Spencertown (15′ Kinderhook quadrangle, N. Y.), the veins resemble a coarse aplite or fine-grained pegmatite. However, in the entire area we have seen no dike or vein to which these names would apply.

Epidote is widely distributed though generally sparse in quartz veins. In a few places, however, the mineral is abundant, and the

FIGURE 7. GLACIATED SURFACE OF RED SLATE, WITH THIN INTERBEDS OF GREENISH SILTSTONE AND SANDSTONE (dashes, dots, and circles), SHOWING INTENSE FRACTURING, AND SMALL DISPLACEMENTS Above northeast-striking faults and quartz veins (right angle dashes). Pasture, ¼ mile northeast of Quackenkill sch...

bright-green veins that penetrate some graywacke outcrops are striking and attract immediate attention (Dale, 1893, p. 39). Epidote is an occasional clastic grain in the graywacke though it also replaces minerals, especially near veins (Pl. 8, fig. 6). The source of lime is difficult to ascertain, but the very low CO_2 in the three analyses of graywacke could be explained by assuming that calcite in the rock has been used up in forming clinozoisite and epidote during the low-grade metamorphism to which the entire terrane has been subjected. Commonly, epidote is associated with little groups of chlorite, but it rarely replaces plagioclase, and albite and epidote are almost mutually exclusive in quartz veins. To what extent these two minerals associate in veins outside the graywacke areas is not known, but the amount of epidote must be very small indeed.

2. *Contiguous Small Graywacke Areas*

Several small areas of graywacke crop out close to the southwestern and southern borders of the upland. Despite their smallness, they are important because any theory of the origin of the upland must also explain the distribution of these small masses. In fact, the difficulties increase as these areas dwindle to individual lenses, barely a few feet thick, amid an argillaceous series in which impure, arenaceous beds are widely distributed.

Loc. 17. Graywacke beds crop out in an elliptical area, a little less than 1 square mile, near *Mud Pond*, about ½ mile south of North Nassau (15′ Troy quadrangle, N. Y.; Dale, 1904a, p. 40; Ruedemann 1930, p. 126; Pl. 1, cross section 28) and 1 mile west of the west border of the graywacke upland. Definite boundaries cannot be drawn, but at several places (hill 820′, ½ mile north of Mud Pond; ridge 820′, ½ mile northwest of Mud Pond) anticlines overturned to the West show greenish and purple silty slate interbedded with typical graywacke. The underlying slate beds are not visible in individual outcrops, but single graywacke beds are at least 6 feet thick. A few coarse zones display quartz and feldspar fragments as much as 9 mm. across.

A large area of graywacke, extending from East Nassau southward to Old Chatham, is shown, incorrectly, on Dale's map (Dale, 1904a, Pl. 1). His report leaves little doubt that he has seen this region only on a few reconnaissance trips. Probably over nine-tenths of the area is underlain by slate and quartzite lenses, but the few places where graywacke crops out are interesting in that they show the close lithologic relations between the graywacke and the argillaceous rocks.

Loc. 18. An abandoned road-metal quarry, on the north side of *hill 900′*, ⅓ mile southwest of East Nassau (extreme southeast corner of Troy quadrangle; Pl. 1, cross section 29), exposes what is believed to be the largest mass of graywacke in this area. The face of the quarry shows a layer of massive graywacke, about 40 feet thick, but subdivided by several silty zones (Pl. 3, fig. 4). As a whole, it dips gently southwest, but is undulating as a flexed and corrugated plate. On the east, north, and west, slate appears to underlie the graywacke, and at the north edge of the quarry a large ledge of purple slate has been left standing in which several graywacke layers, 2 inches thick, dip at 30° S.; at the top of the slate pile, a large mass of graywacke still survives, dipping also at 35° S–SE. North of the quarry, the thick graywacke layer continues and underlies a second hill, crowned by bold ledges of graywacke, and surrounded by interbedded slate. An east-trending axis of a gentle anticlinal fold passes through the north end of the quarry and brings the underlying purple slate relatively high. Although the slate is in places much sheared and squeezed, there is no sign of an unconformity, or thrust plane, that might separate the base of the graywacke from the slate. Oddly, Ruedemann (1930, Figs. 69, 70; p. 151) speaks of a "recumbent fold" in this quarry, and gives two photographs purporting to show it. Neither the photographs nor the plainly exposed, clean ledge surface in the quarry show anything of the sort. The graywacke is an unbroken, nonrepeated sequence of massive layers. Nowhere are there inverted beds or a fold axis. How this error crept into this report is unknown. Conceivably, this reference may have influenced others to assume that elsewhere in the graywacke terrane recumbent folds with flat limbs are present. The writer has not seen a single fold of this kind.

The terminations of the thick graywacke mass are concealed by till, but, at the south base of the quarry hill, a cut at the junction of highways No. 20 and 66 shows a section, 250 feet long, in which purple slate is interbedded with numerous graywacke beds, 1–6 feet thick. On the west side of the cut, the beds dip steeply southwest, they stand perpendicular in the center, and on the east side dips are generally 45°–70° E. There is more slate on this side, and the contortions and intense squeezing of the slate masses around individual graywacke plates are well displayed.

The bed of a small brook draining the west side of this hill exposes also purple slates in which several graywacke beds, a few inches thick, as well as pebble conglomerates can be seen (Fig. 10).

In the area between Brainard and Red Rock, New York, outcrops are poor, but the terrain for the most part is open so that outcrops are rather easily spotted. In the area in which

Dale's map indicates graywacke, the following outcrops have been located:

Loc. 19. Near northern and central *summit 840'*, 1¼ mile south-southwest of Brainard (Kinderhook quadrangle).

Loc. 20. North slope of *hill 1220'*, 1 mile east of Rayville.

Loc. 21. *Hill 640'*, ¼ mile southwest of Riders.

Loc. 22. *Rulland Railroad track*, 1 mile southwest of Brainard.

The rocks are typical graywacke, but only at locality 20 do the layers exceed a few feet in thickness. The surrounding rocks are greenish and purple, silty slates with which the graywacke layers are conformable, as interbeds in a folded, lithologically variable sequence of layers. It would be absurd to assume that each of these thin graywacke lenses is the remnant of a thrust sheet.

The belt of quartzite of Curtis Mountain (Troy quadrangle), Ashley Hill, and west of the Austerlitz graywacke outlier (Kinderhook quadrangle) deserves brief discussion. Near the summit of Curtis Mountain is exposed a series of coarse, feldspathic quartzites, with beds 10–40 feet thick, microscopically very similar to typical graywacke. The only reason for calling it quartzite is that many of the fragments are quartz, and feldspar (both plagioclase and potash feldspar) in the rock is relatively rare. The cement, however, is largely chloritic, and the rock has the greenish color of graywacke. These chloritic quartzites are interbedded with purple and greenish-gray slate.

Though till cover hides the structure of the mountain (Dale, 1904a, p. 22, 23), the same series of quartzitic layers appears to crop out 2 miles south along the strike, at Ashley Hill (Kinderhook quadrangle). Along the north and northwest front of this hill, two or three large, and several less continuous quartzite lenses crop out as low, west-facing scarps. The easternmost quartzite, about 20 feet thick, has numerous zones of slivers and pebbles of dark slate, coarse quartz fragments, and, in one place, its base forms a series of downward-convex troughs that have impinged upon the underlying silty slate and exhibit at their axes a pavement of single, coarse pebbles, 30 to 40 mm. across, overlain by a muddy zone that grades upward into normal, coarse quartzite. In the surrounding rocks abundant albite crystalloblasts were found in a lens of arenaceous slate, or siltstone, about 300 feet west of (below) the northwesternmost quartzite outcrops, and also in places a few feet above this quartzite. Southwest of the summit of Ashley Hill is an important fossil locality.

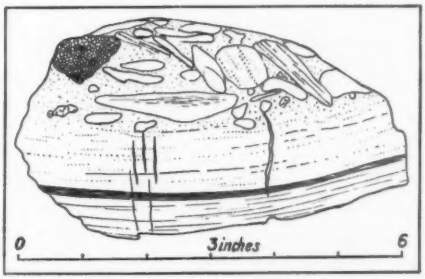

FIGURE 10.—SPECIMEN OF RED SILTY SANDSTONE

Containing a group of pebbles of white, fine-grained sandstone and quartz conglomerate (upper left). Fractures in lower sandstone are filled with quartz veins. West slope of hill 900', ½ mile southwest of East Nassau, Troy quadrangle, N. Y.

Loc. 23. *Lower Cambrian fossil locality, at Ashley Hill* (Kinderhook quadrangle). Originally discovered, it seems, by Foerste or Walcott, it is referred to by Dale (1904a, p. 22–24). The mountain is densely forested, and Dale's exact locality may not have been found. Lower Cambrian fossils were found, however, in a limestone layer, about 1 foot thick, high on the west slope, a little south of the latitude of the summit. On the Kinderhook topographic sheet, the ledge is approximately 1 mm. southwest of the letter "s" of "Ashley Hill". This layer dips 32° W.–SW., not eastward as represented by Dale (1904a, p. 24, Fig. 9), and lies at the bottom of a thicker lens of typical unfossiliferous Schodack-type limestone pebble-conglomerate. East and northwest of the limestone outcrops are quartzite lenses. Where recognizable, bedding dips obliquely east; the rocks suggest a series of folds, with northward-trending axes and asymmetric position of the limbs. The entire series of coarse quartzites is probably Lower Cambrian.

3. Austerlitz Area

Twelve miles south of the southern end of the Rensselaer graywacke upland, another tract of typical graywacke measures 4 miles in a north-south direction, and ¾ to 1½ miles wide. It has been called the Austerlitz outlier by Ruedemann (1930, p. 128; Dale, 1904a, p. 36, 39) as the outcrops are in that township. If the graywacke upland, to the north, is a thrust sheet, this area would presumably represent a smaller

remnant of the same sheet. Ruedemann (1930, p. 128–130) stressed this area in correlating the rock with the Upper Devonian Catskill sand-

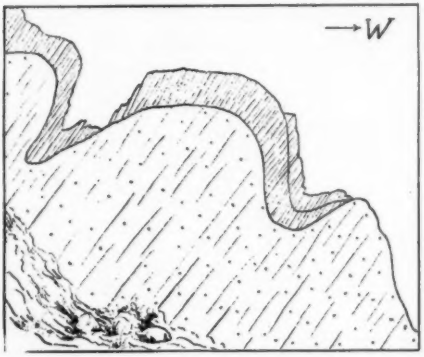

FIGURE 11.—LEDGE OF MASSIVE GRAYWACKE (dotted), INTERBEDDED AND FOLDED WITH GRAY SANDY SLATE

Eastward-dipping slip cleavage obscures bedding in slate, but is hardly recognizable in graywacke. Surface about 2 square yards. Western border of Austerlitz graywacke, 1½ miles south-southwest of Red Rock settlement, Kinderhook quadrangle, N. Y.

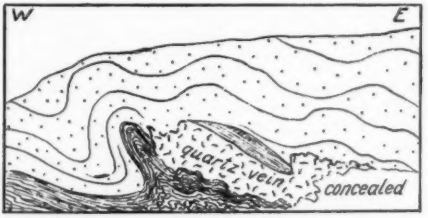

FIGURE 12.—MASSIVE GRAYWACKE, WITH THIN FILMS OF SILT

Showing westward-overturned folds and conformable relation to slate. Length of exposure, 4 feet. South-southwest base of hill 1560′, 1½ miles south of Red Rock settlement, Kinderhook quadrangle, N. Y.

stone. However, it is most difficult to draw a lithologic or stratigraphic boundary of this area. The graywacke layers are underlain, as elsewhere, by silty slates, and on the west side there are conspicuous, impure quartzite lenses interbedded with the argillaceous rocks. The graywacke series is evidently folded in much the same manner as are the surrounding sedimentary rocks. Several outcrops may again be briefly described.

The eastern border is fairly well exposed from State Highway 203 northward to hill 1582′, State Line quadrangle. Greenish-gray slate is the dominant bedrock, and in the outcrops numerous axes of folds trend northeast and plunge either southwest or northeast. A more or less arbitrary boundary has been mapped (Pl. 1), and single lenses of typical graywacke are as much as 300 feet east of this line, folded and conformably interbedded with sandy slate. Northward, the outcrop area narrows, and no graywacke was seen north of Indian Brook (State Line quadrangle), although several lenses of quartzite rise above the till cover north of that stream.

On the west side, the large conglomerate lens, already mentioned, is floored by greenish-gray slate, and a northwestward-flowing brook, ⅛ mile south of Red Rock settlement (Kinderhook quadrangle), exposes a section of silty, gray slate, with many gray and greenish, massive quartzites, 9–15 feet thick, some of which show the same peculiar troughs at the base as were mentioned from Ashley Hill. The quartzites are well exposed southwestward for about 2 miles, dipping eastward with the slates and siltstones, toward the graywacke area. Northeast of Spencertown, the till obscures the details of the western border zone, but the thick conglomerate lens of hill 1360′ rests conformably on black slate of normal easterly dip. At the southern end, graywacke lenses dip northeastward and are conformably folded with greenish slate, near the summit of hill 1400′, ½ mile south of State Highway 203.

Within the area, there are many large ledges of typical coarse graywacke. Most of the rocks are massive, but in a few places they form thick lenses in slate, with axial-plane cleavage dipping about 50° E., and locally imparting a very weak cleavage, or parting, to the graywacke itself (Figs. 11, 12). A most instructive, large cut on State Highway 203 exposes this structure, which is highly misleading in the field for it simulates an elusive bedding in the graywacke (Pl. 5, fig. 1).

Summary of Graywacke Areas

The preceding survey of the graywacke areas has produced no satisfactory evidence of a lithological or structural break between the graywacke and either the underlying or surrounding rocks. Instead, the combined evidence of the many sections, short and incomplete though they are, suggests that a primary change in the type of sediment is recorded in the sense that a dominantly argillaceous series of slates and siltstones received local contributions of relatively coarse detritus, giving rise, first, to greenish-gray, moderately chloritic, locally gritty sandstones (or quartzites). Subsequently, with poorer sorting, typical graywacke lenses were deposited, in which millions of single fragments are embedded in the greenish micaceous paste. Regardless of the direction whence this debris came, the coarsest material is now found near

the western borders of the two largest graywacke areas.

These few facts neither prove nor disprove that the graywacke rests as a thrust sheet on foreign ground, or that a depositional break separates it from other terranes. Such a break must either be sought below the lowermost arenaceous zones or assumed to be everywhere covered by glacial deposits. The first assumption seems the better, for it recognizes the lithologic similarity of the rocks within and below the graywacke areas. The second assumption seems highly speculative because it is most improbable that in so large a contact zone not a single exposure would lie close enough to the thrust to give a hint of its proximity.

The Taconic Area

1. Indian Hill Area

At North Petersburg, the narrow Berlin valley between the eastern edge of the Rensselaer graywacke upland and the Taconic Range widens. Hoosic River flows northward, but the crest of the Taconic Range veers northeastward, and between the river and the receding crest of the range lies a group of low hills, culminating in Indian Hill (North Pownal quadrangle, N.Y.-Vt.; Pl. 1). Although small, the Indian Hill area has general stratigraphic interest for the lithology and association of rock types are typical of the so-called valley sequence of slate and limestone described in considerable detail in regions farther north and northwest. It would have been convenient to divide the description into a stratigraphic and a structural part, but the generally small and discontinuous outcrops make it impossible to establish a reliable geologic sequence.

Within the map area, the Indian Hill district consists of a central belt of gray and black slate, whose strike makes an arc convex to the northwest, and a northwestern and a southeastern limestone belt that conform to this arcuate form. Bedding, well preserved in most outcrops, dips nearly everywhere southeast or east, and fossils of Beekmantown age have been found in a small railroad cut, 1 mile south of Hoosick (Prindle and Knopf, 1932, p. 273; Pl. 1), in rocks that belong very probably to the northwestern limestone belt. Just north of the map area, L. D. Bonham (1950) located Beekmantown fossils in limestone that appears to be continuous with the southeastern limestone tract (at Whipstock Hill, 7½′ Hoosick Falls quadrangle). The two carbonate belts seem to join below glacial deposits, 1½ miles north-northeast of North Petersburg.

At least three small, discontinuous limestone tracts are exposed within the central slate zone, dipping also southeast. They may be calcareous lenses in the slate, or their contacts may be due to strike faults. The southwesternmost limestone narrows from 60 feet, at the northeast end, to about 5 feet at the southwest end where both the top and bottom contacts are exposed in a road cut. They look like primary contacts, but one of them could be a smooth strike fault that has cut out part of the originally thicker limestone.

Breese Hollow, the depression that marks the southeastern limestone tract, is so heavily covered by till that the contacts of that limestone with the Indian Hill slate are concealed. Presumably a high-angle thrust separates the two, but possibly there is a primary sequence, limestone-slate-limestone, perhaps somewhat disturbed by subsequent faulting. The southwestern continuation of the three belts is terminated at North Petersburg, probably by a north-south striking high-angle fault along which the eastern block has been dropped. The reasons for this assumption will be given later.

On the southeast, the Breese Hollow limestone tract approaches the Taconic Range. The discussion of that zone will be resumed after the rocks in the Berlin valley have been described.

2. Berlin Valley

General statement.—The meridional valley that separates the eastern border of the graywacke upland from the Taconic Range is about 20 miles long. Its northern portion is narrow, with fairly steep, forested slopes, but south of Center Berlin it widens as the front of the graywacke upland recedes southwestward from the straight front of the Taconic Range. Black slate and limestone, probably Ordovician, and perhaps correlating with the rocks of the Indian Hill area, underlie the southern valley. The nar-

row, northern valley is incised into greenish-gray siliceous slates which, at least in one place, have yielded Lower Cambrian fossils. The geologic features are described from north to south.

Area North Petersburg-Berlin.—The western slopes of the valley between North Petersburg and Berlin are underlain by the greenish-gray slates, with interbedded siltstone beds, and impure quartzite lenses that have been mentioned in the description of the eastern border of the graywacke upland. Several lenses of tuffaceous slate, in conformable contact with the ordinary types, have been marked on the map (Pl. 1). At a low bluff, about 400 feet west of State highway 22, half a mile southwest of Petersburg, the slate contains pebbles of fine-grained white, lavender, and greenish quartzose rocks, 1–5 inches in size, strongly deformed and drawn out into spindles, and this may be a small lens of pyroclastic material.

Northwest of North Petersburg, the east-west striking argillaceous rocks form a bold scarp facing east. At its east base, due west of the village, a mass of limestone dips gently west. Talus conceals the contact, but the writer excavated and found a small portion of the contact, near the northwest end of the limestone. It appears to be an almost vertical fault, and the rocks in immediate contact are broken. One mile south, another mass of limestone is exposed on State Highway 22. The beds are nearly horizontal, and in the weathered surfaces (since destroyed by blasting) the minute details of stratification as well as of intraformational folds and corrugations are beautifully visible. The hillside directly west of this mass is entirely slate, which shows small-scale folds and slip cleavage. The anomalous juxtaposition of the two rocks can be explained best by a high-angle fault (Pl. 1). As the slate, but not the limestone, appears to have been deformed, the eastern block has probably dropped along this fault; this is also the most probable explanation for the northern fault. Possibly these two faults are parts of the same fracture, which may extend both south and north beyond the area mapped.

South of the limestone blocks, the valley is crossed by a thick belt of the greenish, siliceous slates, striking southeast. The internal structure of this belt is obscure. Many small-scale folds can be seen in outcrops 1–2 miles north-northwest of Petersburg, but slip movements along cleavage planes (Figs. 13–15) have also accommodated the deformation of the beds. In a large, fresh cut of State Highway 22, 2 miles north of Petersburg, the only folds seen are the minute corrugations common to nearly all the argillaceous rocks of the region.

Where this belt of slates crosses to the east bank of Little Hoosic River, a zone of strong, impure quartzite lenses forms the steep western slopes of Odell, Moon, and Poplar Hill. At the base of Moon Hill, amid the hundreds of barren ledges obscured by forest and dense underbrush, Prindle located a few thin limestone

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PLATE 5.—CLEAVAGE IN GRAYWACKE AND SLATE

FIGURE 1.—MASSIVE, FLAT-LYING BED OF GRAYWACKE, RESTING ON THIN SILTSTONE BED WITH QUARTZ-ALBITE VEIN (BELOW CENTER)

Crude slip cleavage dips 55° E. Cut on State Highway 203, 1½ miles east of Spencertown, New York. Looking north.

FIGURE 2.—GREENISH-GRAY, GLOSSY SLATE WITH THIN SILTSTONE BEDS, DIPPING 10° E., CROSSED BY SLIP CLEAVAGE DIPPING 45° E.

In cut of State Highway 2, at west entrance to Petersburg, New York. Looking north.

FIGURE 3.—BLACK SLATE WITH GRAY INTERBEDS, PROBABLY ORDOVICIAN

Showing westward-overturned folds, crossed by axial-plane slip cleavage dipping 40° E. Anticline on left side has been dropped 1 foot by vertical fault. Southwest spur of Butternut Hill, Hancock quadrangle, Mass.-N. Y.

FIGURE 4.—FLAT-LYING ?ORDOVICIAN BLACK SLATE

With feeble fracture cleavage (front to back), striking north-south, and vertical east-west joints. Looking south. South bank of brook, ½ mile north of Sacred Heart Cemetery, at Stephentown Center, New York.

FIGURE 2

FIGURE 4

FIGURE 1

FIGURE 3

CLEAVAGE IN GRAYWACKE AND SLATE

FIGURE 2

FIGURE 4

FIGURE 1

FIGURE 3

DEFORMATION OF LIMESTONE AND SLATE NEAR THRUSTS

lenses, ½–2 inches thick. After extraordinarily patient and tedious collecting, he found identifiable Lower Cambrian fossils in a few chips. Every geologist who knows the difficulties of finding fossils in this densely forested terrain will pay the highest tribute to Prindle for this contribution to our knowledge because it proves that at least a part of this greenish, siliceous slate series is Lower Cambrian. At the United States National Museum, a few additional chips, collected by Prindle, bear the inscription, "½ mile northwest of Petersburg", and refer undoubtedly to hill 1220′, where ledges of the same siliceous slate are exposed. The writer has not seen any limestone lenses there, but he located two additional exposures where blasting may some day yield fossils. One is a small ledge at the intersection of State Highway 22 and the Rutland Railroad, at the northern entrance to Petersburg; the other is at the extreme west base of hill 1360′, half a mile northwest of Prosser Hollow (North Pownal quadrangle).

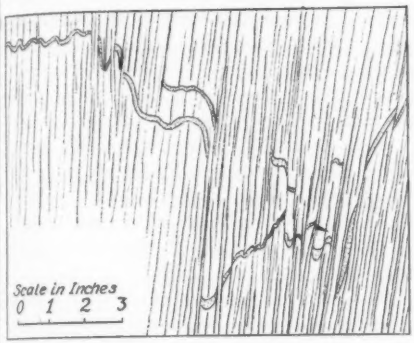


FIGURE 13.—DISPLACEMENTS OF THIN SILTSTONE BEDS IN GREENISH-GRAY, SILICEOUS SLATE

Caused by dissection of the rock by slip-cleavage planes. Looking southeast. Summit of hill 1220′, ¼ mile north-northwest of Petersburg, North Pownal quadrangle, N. Y.-Vt.

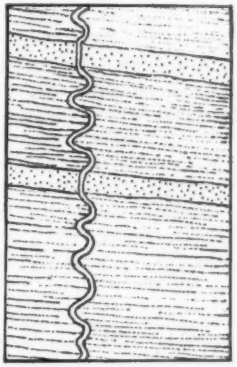


FIGURE 14.—VERTICAL QUARTZ VEIN, TRANSECTING GENTLY EASTWARD-DIPPING LAYERS OF SLATE (parallel lines) AND SILTSTONE (dots)

300 feet northwest of locality of Fig. 13. The vein is straight where it cuts siltstone, but folded where it passes through slate, probably due to slip parallel to bedding in the slate. Height of exposure, 2 inches.

PLATE 6.—DEFORMATION OF LIMESTONE AND SLATE NEAR THRUSTS

FIGURE 1.—?ORDOVICIAN LIMESTONE, STRONGLY DEFORMED

With *a* lineation plunging with dip of *s* planes at 30° E.-SE., parallel to hammer handle. Looking northwest, in pasture, ¼ mile southeast of Hancock, Mass.

FIGURE 2.—TYPICAL STRUCTURE OF CHLORITOID-BEARING QUARTZ-CHLORITE SCHIST OF THE TACONIC RANGE

In railroad cut south of North Pownal station, Vermont. Looking east-northeast. Open folds of moderate wave length, overturned to the west, crossed by east-dipping axial-plane cleavage, and *b* lineation parallel to subhorizontal axes of folds.

FIGURE 3.—LOOKING EASTWARD AT DEFORMED QUARTZ-CHLORITE SCHIST OF A THRUST ZONE

In railroad cut, at North Pownal station, Vermont. *s* planes dip east, coinciding locally with bedding planes. A stronger bed forms the right edge of the ledge, showing axis of fold coinciding with dip of thrust zone. Thin white lines in left part of surface are quartz veinlets folded about axes in identical direction. Main rock surface is a cross joint, dipping west-northwest, perpendicular to *a* lineation.

FIGURE 4.—CONTACT, AT HEAD OF HAMMER, OF QUARTZ-CHLORITE SCHIST AND LIMESTONE

Probably Ordovician, 600 feet east of U. S. Highway 7, 1½ miles southeast of Pownal. Contact suggests undisturbed superposition of schist on limestone.

Most of the thin limestone layers crop out under the water of Little Hoosic River, making excavation difficult, but lithologically all three localities seem to belong together.

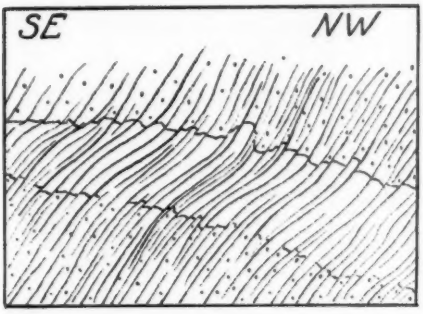


FIGURE 15.—FLAT GRAYWACKE LAYERS (dotted), INTERBEDDED WITH GRAY SLATE

Showing small displacements along slip-cleavage planes that dissect both rocks. Length of surface, 2 inches. One hundred and fifty feet northeast of summit 1612′, 1½ miles northeast of Taconic Lake, Taborton quadrangle, N. Y.

The structure of the rocks southeast of North Petersburg will be discussed in connection with the description of the Taconic Range.

Within the valley proper, there are probably several local fault zones. At the west entrance of Petersburg, for instance, State Highway 22 exposes a ledge in which the green slates lie nearly flat and are crossed by eastward-dipping slip cleavage (Pl. 5, fig. 2); but the same rock, 1000 feet southeast, has been sheared and folded, quartzite and graywacke lenses have been ruptured, and the fragments separated. Six hundred feet east, in the bed of Little Hoosic River, the siliceous greenish slates dip east, show hardly any folds, and even cleavage is restricted to a few thin, argillaceous beds. Unfortunately, extensive alluvium and till in the valley bottom obscure the structural details of the rocks. Prindle and Knopf (1932, Fig. 2) report outcrops of Ordovician black slate and small masses of limestone a short distance south of Petersburg. These have not been found, but a few small outcrops are easily overlooked in the forested slopes. The greenish-gray slates, with thin lenses of graywacke and argillaceous quartzite, underlie the higher slopes of the valley, both north and northeast of Berlin, and as a whole the series seems to dip eastward although probably complicated in detail by small folds and faults.

Area Berlin-Stephentown.—South of Berlin, the steep slopes of greenish siliceous slates that hemmed in the valley north of the village recede, and in the widening lowland appear low ridges and hills, underlain by black slates with associated gray limestone masses, unlike any of the rocks to the west, north, or east.

BLACK SLATE: The rocks are probably of Ordovician age (Prindle and Knopf, 1932, Fig. 2). Though they appear to be important for the interpretation of the structure and age relations in this area, these rocks do not seem to have been described.

One of the best outcrops of the black slate is a northwest-trending ridge that starts ¼ mile west of Cherryplain (7½′ Berlin quadrangle, N.Y.-Mass.-Vt.; Pl. 1) and extends just beyond the eastern border of the Taborton quadrangle, to the steep southeast slope of Rocky Hill. The ridge is composed of a plate of slate that dips about 25° NE. but is thrown into millions of minute crinkles and crenulations whose axes trend and plunge northeast, very nearly parallel to the dip of the planes. Larger folds or faults seem to be absent from this ridge, and there must be about 400–500 feet of slate in this section. In contrast to practically all other argillaceous rocks in the area, this black slate has but a few, very thin, gray or greenish siliceous interbeds. Thicker quartzite, siltstone, or graywacke beds have not been seen.

This ridge of black slate reaches an abrupt northwestern end at a col, at an elevation of 1500 feet, ½ mile southeast of the summit of Rocky Hill (eastern border of Taborton quadrangle). The higher portions of the mountain consist of a great mass of complexly folded greenish-gray siliceous slates so unlike the black ones, and with strikes, dips, and orientation of fold elements consistently in such divers directions, that it is very probably a different series of rocks. At the col the two formations approach to within 30 feet of each other; presumably a high-angle fault separates the two, and has dropped the black slate down on the east side. The matter can not be settled definitely, but the evidence is favorable.

The slate, with the same characteristic "washboard" crenulations, is well exposed on the east side of the valley between Berlin and Center Berlin, and for several miles south of that settlement. Small but excellent exposures are on, and directly east of, State Highway 22, south-southwest of the top of Butternut Hill (Hancock quadrangle). Here too the layers of deposition display the small zigzags and contortions, the axes plunge northeast, and in one place the beds show an isoclinal, westward-overturned fold crossed in textbook fashion by axial-plane slip cleavage (Pl. 5, fig. 3). Isoclinal folds, at least on a small scale, are seen occasionally in the greenish-gray, siliceous slates that surround the graywacke upland. In the black slate they seem the exception rather than the rule.

Two miles farther south (in the beds of several eastward-flowing brooks, ½ mile north of Sacred Heart Cemetery at Stephentown, Stephentown Center quadrangle), the layers of deposition of the slate are horizontal or very slightly warped by widely spaced fracture-cleavage planes that strike north-northeast (Pl. 5, fig. 4; Pl. 1, cross section 15). Evidently the rock has not been deformed to any appreciable extent here, and it is most unfortunate that till conceals the contact between this flat-lying black slate and the Rensselaer graywacke. The southernmost exposures of the slate visited, are on West Hill, a large rise south of Stephentown (Stephentown Center quadrangle). Here the slate has been deformed much more than in the Berlin valley and in an abandoned quarry $\frac{3}{10}$ mile south-southeast of Presbyterian Hill Cemetery, on the west side of a local road, is one of the few truly recumbent folds, with instructive displacements along oblique and vertical, east-west striking faults and fractures (Dale, 1893, p. 320, 321; Pl. CI, fig. 26). The quarry is obsolete, but the fold is still visible, and on the dump one can collect most interesting slabs showing the displacements in the slate, parallel to a well-developed *a* lineation.

LIMESTONE MASSES BETWEEN BERLIN AND HANCOCK: Between Berlin and Hancock, limestone masses appear in the belt of black slate. One of them has furnished Ordovician fossils; probably the others are also of Ordovician, perhaps Trentonian age. The contacts of several are interesting and well exposed. They are briefly outlined from north to south.

LIMESTONE MASS SOUTH OF BERLIN: One mile south-southwest of Berlin, a complex of hills rises west of Little Hoosic River, culminating in hill 1530′ (Taborton quadrangle). This rise, as well as several others south of it, are underlain by a large mass of limestone and dolomite, set off from the graywacke upland by a depression. Toward the river, the hills drop off in steep, forested slopes. Details of the boundaries, as far as known, and internal structures of the rocks are shown on Figure 16, and cross sections 9 and 10 of Pl. 1. No contacts of the surrounding argillaceous rocks are exposed, but, within the limestone mass, tracts of fine-grained, black slate come to the surface, with local interbeds of gray, arenaceous slate, siltstone, and quartzite. Associated with the largest slate zone, near the center, are graywacke lenses of which one, west of summit 1530 , is indistinguishable from typical Rensselaer graywacke. It is tentatively assumed that this is a fault block that dips obliquely southeast, under the limestone, though it could be explained in other ways also.

Lithologically, the limestone resembles other Ordovician carbonate rocks of the area. Microscopically, some clastic quartz and feldspar is found in most specimens, and in the walls of several obsolete quarries, near summit 1530′, cream-colored blocks of massive dolomite, fractured and pulled apart, are surrounded by thinly laminated calcite limestone that has been kneaded into the spaces between the dolomite blocks. Elsewhere, the limestone displays tight isoclinal folds and recrystallized shear zones, and yet only 1000 feet away, at the western summit of hill 1530′, the original beds of deposition are beautifully preserved, in subhorizontal attitude, with scores of thin sandstone layers and films of silt interbedded with arenaceous limestone, the whole series cut by a few slip planes and a crude, east-dipping slip cleavage. Both normal and reverse movements along these planes, with displacements rarely exceeding an inch, are common and stronger sandy beds are dragged where they are traversed by the slip zones. The mass of Rensselaer graywacke underlying this part of the limestone mass a few hundred feet away makes one wonder whether the graywacke could be the original floor upon which this limestone, impure and arenaceous as it is, was deposited. As so often in New England geology, the exposures are not continuous enough to decide the matter, but the possibility should be considered.

Limestone approaches the eastern border of the graywacke upland to within about 300 feet at the small double hilltop, 1670′, ⅓ mile east of hill

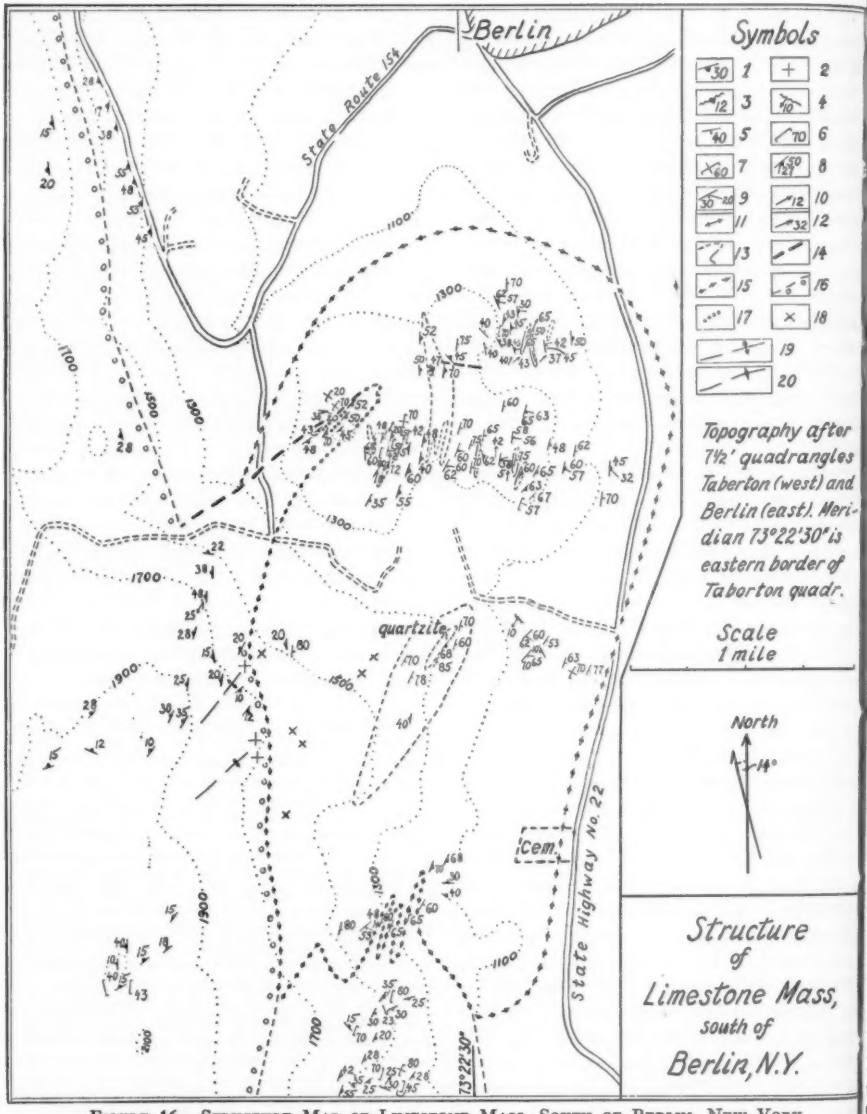


FIGURE 16.—STRUCTURE MAP OF LIMESTONE MASS, SOUTH OF BERLIN, NEW YORK

Explanation of Symbols:

1. Strike and dip of beds.—2. Horizontal beds.—3. General strike and dip of contorted beds.—4. As in 3, with bearing and plunge of axes of folds.—5. Strike and dip of *s*-planes.—6. Strike and dip of slip cleavage.—7. Beds horizontal, slip cleavage dips 60° SE.—8. Bedding as in 1, with trend and plunge of lineation.—9. *s*-planes as in 5, with trend and plunge of lineation.—10. Trend and plunge of axis of fold.—11. Horizontal axis of fold.—12. Trend and plunge of lineation and axes of folds.—13. Contacts of slate zone in limestone; solid line where exposed, dash line where concealed.—14. Inferred fault.—15. Outer contact of limestone, largely conjectural on north and northeast.—16. Western border of Rensselaer graywacke upland.—17. Graywacke in slate lens.—18. Isolated rock outcrops.—19. Axis of inferred anticline.—20. Axis of inferred syncline.

2064′, 1.1 mile southeast of Kendall Pond (Taborton quadrangle). Both the basal beds in the graywacke scarp and the westernmost limestone outcrops dip at about 20° W., but till covers the interval.

The southernmost limestone outcrops are in the bed of an east-flowing brook, ½ mile southwest of a cemetery near Center Berlin. The limestone is conformably folded with several lenses of black slate, whereas, about 1000 feet south, a complex of greenish siliceous slates makes its appearance, so different in all respects from the black slate that the writer regards them as a different rock series; but again till obscures the contact.

The writer believes the black slate represents argillaceous deposits of essentially the same age as the limestone, whereas the surrounding greenish slates, graywackes, and similar arenaceous rocks may be parts of an older terrane upon which the Ordovician rocks were deposited. This floor would have been warped subsequently, and further deformation must have affected both formations producing the present, fairly complicated relations. On the west, the limestone complex may be set off from the graywacke upland by a normal fault, and possibly another normal fault lies east of the limestone, dropping the eastern block. This has been assumed in cross sections 9 and 10 of Plate 1, though the evidence is not compelling.

LIMESTONE ZONE SOUTHEAST OF CENTER BERLIN: Hill 1580′, $\frac{3}{10}$ mile due east of Center Berlin (7½′ Berlin quadrangle), exposes a plate of limestone about 100 feet thick that dips about 40° E. and can be followed along the strike for about a quarter of a mile north and a half a mile south. The limestone rests on crinkled, black slate, but in places also on greenish, chloritic, and arenaceous slate, and is terminated on the east by a thrust fault zone that dips at about 40° E.

Structural criteria for thrust zones: In this poorly exposed region, where many conclusions must rest on inference, it is important to list a few reliable structural criteria which indicate thrusts. In Dutchess County, New York, the author (Balk, 1936, p. 737–745) could show that thrust faults are accompanied by a narrow plate of rocks showing directions of deformation not found elsewhere. The Paleozoic sedimentary rocks of that region strike about north-south, and folds are overturned to the west. These folds are traversed by axial-plane slip cleavage (in the earlier paper called fracture cleavage, but the term slip cleavage is preferable for most of the argillaceous rocks), and the lines of intersection of folded beds and cleavage produce a north-south lineation of low plunge which, according to common terminology, is called a *b* lineation. In the vicinity of thrust faults, however, the slip-cleavage planes have functioned as planes of intensified slip. A lineation parallel to the movements penetrates the entire rock fabric and is most intense along the principal thrust surface except where modified by subsequent silicification or other mineralization. As the lineation parallels the tectonic transport, it is referred to as *a* lineation. Close to thrust faults, the transformation of the structure can be followed through individual stages, and many illustrations of it have been given in the 1936 paper. In the northern Appalachian area, almost all the thrusts dip east, and the *a* lineation plunges east or deviates a little to the east-southeast. In the most intensely deformed zones, one finds occasionally folds whose axes plunge also east, parallel to the *a* lineation (Pl. 6, fig. 3). The crystallographic *c* axes of quartz in the laminated zones close to thrust faults arrange themselves preferentially in a plane perpendicular to the *a* lineation (so-called *bc* girdle), and this has been verified in a study of the fabric of the Lower Cambrian Cheshire quartzite in this area (Balk, 1952). Joint systems are also associated with the thrust stucture so that each lineation has a fracture system oriented perpendicular to it. With the regional *b* lineation is associated an east-west fracture system of steep dips, whereas the *a* lineation is crossed by north-south trending joints that dip westward at angles complementary to the plunge of the *a* lineation. Plate 10 illustrates the manner in which these two lineations and associated structures can penetrate a complexly deformed mass of rocks. In the writer's experience, eastward-plunging lineation in this region always indicates some thrusting, and, if the appropriate symbols in Plate 1 are located, there are probably many times more thrust zones in the region than have been mapped. But the quality of exposures, and the map scale impose limitations. Cloos (1947) has given an illuminating account of the mechanics of deformation that produces *a* lineation in the central Appalachians, and many other papers (Cloos, 1946; Fellows, 1943; Kvale, 1945) discuss the same problems.

Along this thrust fault, green quartz-chlorite schist of the western Taconic Range has overridden the limestone. Near it, the limestone is locally sheared to a finely laminated calcareous schist, with strong *a* lineation plunging at about 40° E-SE. In places, thin lenses of the chlorite schist appear within the most intensely sheared limestone, and narrow tracts of black slate are also incorporated here and there in the deformed zone which is about 20–100 feet thick.

BUTTERNUT HILL LIMESTONE MASS: Two small limestone lenses, about half a mile southeast of Cherryplain (7½′ Berlin quadr., N.Y.-Mass.; Pl. 1), indicate a connection of the limestone just described with a larger mass farther south that underlies Butternut Hill (northwest corner of Hancock quadrangle, N.Y.-Mass.; Pl. 7). It rises as a bold cliff above the smooth till cover of a low ridge east of the valley, and the summit is underlain by a mass of compact dolomitic limestone that drops off sharply at the west and south. At the west base, the massive rock is floored by thin-bedded, laminated limestone dipping at about 40° E., and this in turn rests

on black, graphitic slate, in every respect like the slate elsewhere in this valley.

A short distance north of the summit this sequence is overlain by till, but on the south side limestone continues for about a mile. The massive, dolomitic block ends about half a mile south, and the flooring limestone sustains the hill slope farther south, displaying a gently undulating surface of bedding planes that slope about 10° W. A small, abandoned quarry exposes beautifully details of the primary stratification, and it is probably from here that Walcott (1888, p. 239) obtained a small collection of fossils of Ordovician, probably Trentonian, age. The specimens, deposited at the United States National Museum, are not good enough for specific identification, but the lithology and the general aspect of the fauna resemble closely other Ordovician collections from the Taconic area. Near the southwest end of the exposures, the limestone is interbedded with the black shale, and bedding planes and axes of folds slope eastward and northeastward, into the limestone.

The southeastern end of the limestone is a shallow syncline, plunging about 5° N., and here, too, the limestone is floored by the black slate. However, the outcrop of limestone narrows northward, and directly east of it green quartz-chlorite schist appears as a ridge which can be followed all the way to the top ledges of Butternut Hill. The contact of the two rocks is repeatedly exposed. The surface dips at 40° E., and the well-developed *a* lineation in both rocks, combined with strong lamination, clearly marks the surface as a thrust. Directly east of the top of Butternut Hill, a narrow plate of quartz-chlorite schist branches off the main mass, and a wedge of limestone separates it from the main schist plate for about 300 feet (Pl. 7, cross section 3).

Directly in the southern extension of this western thrust plate of schist, the well-exposed bedding planes in the limestone form the crest of a gentle anticline, suggesting that the schist continues underneath, and that its rise has arched the superimposed layers. Cross section 4 of Plate 7 gives the writer's interpretation of the relationships. At the northern boundary of the limestone, another low anticlinal axis, trending west-northwest, may have formed in a similar way by the movement of the underlying schist.

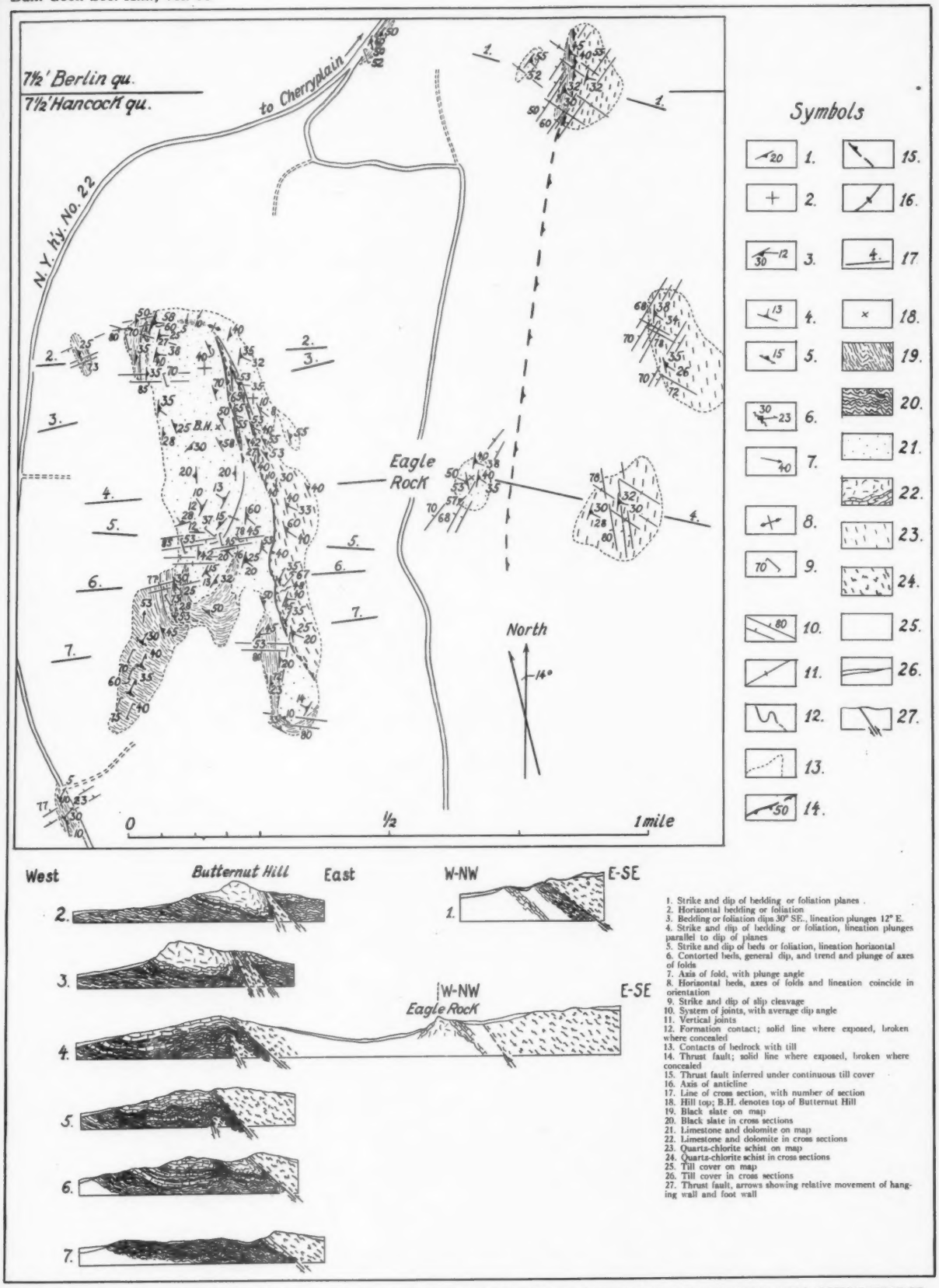
Two hundred feet north of the north end of the schist, the thin-bedded calcitic limestone that appears to underlie the dolomite mass of the hill top continues like a horizontal platform that skirts the northeast flank of this remarkable hill. Unfortunately, the contact between this limestone plate and the schist is concealed. The writer suspects it is a fault along which the schist has broken through the limestone (Pl. 7, cross section 2).

A quarter of a mile east of Butternut Hill stands Eagle Rock, a small mass of dolomite and laminated calcite limestone. Till conceals its contacts but the calcitic rock shows strong lineation bearing east-southeast on variously dipping surfaces, as does also a mass of schist 800 feet east.

Perhaps the black slate and limestone represent a plate of Ordovician sediments that was deposited on an underlying schist formation which has been thrust-faulted so that it penetrated and folded the overlying rocks. At least two thrust faults are required, and there may well be additional ones, partly concealed. The eastern thrust, east of Eagle Rock, is actually exposed ¾ mile north of the rock. Black slate is here overlain by the quartz-chlorite schist and limestone lenses are directly west. Here, too, all rocks show the strong *a* lineation, are intensely sheared, and locally silicified (Pl. 7, cross section 1).

LIMESTONE NEAR HANCOCK: The trace of the eastern thrust fault just mentioned can be followed southward to the vicinity of Hancock, Massachusetts (Pl. 1). A larger limestone mass forms hill 1333′, ½ mile northeast of Stephentown. Although its eastern contact is not exposed, the rock at the west base has been deformed into a thoroughly schistose rock, with *a* lineation in the regional east-southeast direction grooving the surfaces of every slab.

Southeast of Hancock, additional limestone ledges attest equally intense deformation of the carbonate rocks. Instructive ledges are on the east side of Kinderhook Creek, north of Brodie Mountain road (7½′ Hancock quadrangle, Mass.-N.Y.), and also in a pasture, ¼ mile east-southeast of the village of Hancock, where the *a* lineation is extraordinarily well developed.



1. Strike and dip of bedding or foliation planes.
2. Horizontal bedding or foliation
3. Bedding or foliation dips 30° SE., lineation plunges 12° E.
4. Strike and dip of bedding or foliation, lineation plunges parallel to dip of planes
5. Strike and dip of beds or foliation, lineation horizontal
6. Contorted beds, general dip, and trend and plunge of axes of folds
7. Axis of fold, with plunge angle
8. Horizontal beds, axes of folds and lineation coincide in orientation
9. Strike and dip of slip cleavage
10. System of joints, with average dip angle
11. Vertical joints
12. Formation contact; solid line where exposed, broken where concealed
13. Contacts of bedrock with till
14. Thrust fault; solid line where exposed, broken where concealed
15. Thrust fault inferred under continuous till cover
16. Axis of anticline
17. Line of cross section, with number of section
18. Hill top; B.H. denotes top of Butternut Hill
19. Black slate on map
20. Black slate in cross sections
21. Limestone and dolomite on map
22. Limestone and dolomite in cross sections
23. Quartz-chlorite schist on map
24. Quartz-chlorite schist in cross sections
25. Till cover on map
26. Till cover in cross sections
27. Thrust fault, arrows showing relative movement of hanging wall and foot wall

STRUCTURE MAP OF THE VICINITY OF BUTTERNUT HILL, SOUTH OF CHERRYPLAIN, NEW YORK

and the axes of folds plunge east-southeast, parallel to it (Pl. 6, fig. 1). The distribution of the limestone in this particular area is complicated. Another thrust zone entering from the northeast is assumed on the basis of exposures of quartz-chlorite schist and deformed graywacke, east of the limestone.

3. *Taconic Range*

Lithologic characteristics.—Rocks of the Taconic Range have been collected from Mount Anthony southward for about 26 miles to the vicinity of Hancock, Massachusetts. Previously no detailed studies had been made of these rocks, and it was thought that a number of traverses, supplemented by examination of many specimens and thin sections, would bring to light a greater variety of lithologic types than appears from the scanty references in the literature. However, the lithology of the rocks, as a group, is indeed monotonous. Most abundant are greenish-gray, rarely purple, glossy, fine-grained and well-laminated chloritic schists, so little metamorphosed that the terms slate, phyllitic slate, or metashale (Kesler, 1950, p. 8) are equally applicable. But as the term schist is commonly applied in the English literature to chloritic low-grade metamorphosed argillaceous rocks, it seems best, for the present purpose, to use the designation quartz-chlorite schist. Interbedded with the thinly laminated schists are heavy-bedded, more quartzose rocks that may be called siltstone. Plate 1 indicates a number of lenses of greenish graywacke and conglomerate.

In the upper part of the western slopes of the range, the greenish rocks show minute, dark-greenish dots, which are crystals of chloritoid. On the basis of thin sections from about 80 specimens, a chloritoid isograd has been drawn (Pl. 1) but, apart from a few sporadic showings of albite, no evidence was seen of other minerals that would suggest higher metamorphic grades in the Taconic Range proper.

Description of Rock Types:—Common chlorite schist of the range consists of a well-laminated aggregate of thin lenses of granular quartz and sericitic muscovite, with variable amounts of weakly pleochroic chlorite. Many specimens have been studied to try to determine the general character of the chlorite, but the small grain size (about .01 mm) and the intimate intermingling with sericite made satisfactory determinations impossible. From the vicinity of a quartz vein, on hill 1820′, ½ mile south of Shimwell Hollow (7½′ Berlin quadrangle, N.Y.-Mass.), the writer isolated a light-green, weakly pleochroic prochlorite ($\alpha = 1.634$), and perhaps the examination of relatively coarse phases of the schist may result in a better understanding of the composition of this mineral. A few slides have two chlorites differing in pleochroism, and differences in interference colors are common. Some flakes show the deep berlin blue, or a reddish brown, instead of the more common dark gray.

Chloritoid occurs as groups of small (.1–.25 mm.) plates, exhibiting the common pleochroism of that mineral, but in light shades. Twinning and other optical data are substantially the same as those of chloritoid from Dutchess County (Barth and Balk, 1934). The crystals may lie singly in random directions in the schistose rock, or may form little groups of radiating crystals. Larger chloritoids enclose the dark dust films of the schist groundmass as internal *s* (Pl. 9, figs. 1, 2). Specimens from the western slopes, east of Center Berlin, show the original bedding planes of the schist folded and disrupted, and the surfaces of these fragments have served as planes of attachment for the chloritoid crystals which, with secondary quartz aggregates, fill the interstices. On the other hand, chloritoid crystals in schist at the railroad cut at North Pownal, Vermont, are older than the thrust faults and have been deformed by slip planes. In intensely deformed schist, the mineral is sparse, and in its stead are found here and there ellipsoidal metacrysts of chlorite, about .1 mm. across. Table 1, No. 3 gives a chemical analysis of chloritoid schist.

Lenses of graywacke, interbedded with the quartz-chlorite schist, have been marked on Plate 1. Their mineral composition is like that of the surrounding rocks, and apart from their more pronounced schistosity and coarser groundmass fabric which commonly replaces clastic fragments (Pl. 9, figs. 3, 4), they are lithologically similar to the main graywacke area. Table 1, No. 2, gives the chemical composition of a lens, east of Center Berlin. The writer doubts that there are in the Taconic Range continuous areas of graywacke as large as shown on a preliminary map by Prindle and Knopf (1933, Pl. 14). The rocks are all lenses, rarely thicker than 10–15 feet, and some are only a few inches thick, yet may be coarse, even conglomeratic. A good example is on the east base of Poplar Hill, 1.3 miles east-northeast of Petersburg. The layer is only an inch thick, but is studded with single quartz fragments, 3–8 mm. in diameter, and fewer of feldspar, embedded in the characteristic green groundmass of quartz, sericite, and chlorite.

Loc. 24. Prindle and Knopf (1932, p. 290) mention a coarse, black sandstone, or conglomerate, northeast of Stephentown. This refers probably to outcrops in a ravine, due west of summit 2458′ 3¼ miles northeast of the village where such a rock is interbedded with gray and greenish siltstone and slate. The dark color is due to abundant graphite which pervades not only the quartz-mica groundmass of the rock but, oddly enough, many quartz grains as well, as groups of exceedingly minute parallel films (Pl. 9, figs. 5, 6). It throws a most interesting light on the behavior of graphite during the folding and thrusting of the rocks under conditions of low metamorphism, and the films may

have fixed rupture planes inside the quartz grains that would otherwise perhaps be recognized as deformation lamellae. Within each quartz grain, the films of graphite follow a uniform direction.

A peculiar rock was found in loose boulders in the bed of a brook that drains the west slope of Berlin Mountain (7½′ Berlin quadrangle). Thick shells of black iron or manganese oxides, when removed, reveal a pink and green, mottled rock, replete with fragments that resemble megafossils, over 10 mm. in length. If they are of organic origin, they have lost all structure. Those of the pink variety are largely ankeritic carbonates, whereas, in the dark-green rocks, a fine-grained groundmass of quartz, some feldspar, and strongly pleochroic chlorite, with admixtures of iron oxide, is traversed by carbonate veinlets that have plainly replaced the groundmass and project into it with numerous edges of rhombohedra (Pl. 11, fig. 3). Unfortunately, the outcrops of these boulders could not be located.

Fairly common in the schists of the range are gray pellets, composed of fine-grained quartz and sericite, but without chlorite. They lie singly in greenish or purple schist, or occur in such numbers that one is reminded of sheared pyroclastics or pebble conglomerates. They may be seen through a considerable thickness of gray siltstone in the bed of the brook that drains the southwest slope of Misery Mountain (7½′ Hancock quadrangle, Mass.-N.Y.). The pellets are deformed into triaxial ellipsoids; common dimensions are: 1/16 to ¼ inch perpendicular to the local *s* plane, ¼ to ½ inch parallel to the strike of the *s* plane, and 2 to 3 inches parallel to the dip of the *s* plane. Many pellets that have weathered out of the ledges have fractured along cross joints, and their total length may have been considerable.

In the Taconic Range there is said to be a gradation from unmetamorphosed to albite-bearing schists, from west to east. The writer has examined 142 thin sections from various locations in the range. The only rock in which albite porphyroblasts are conspicuous is the tuff belt, already mentioned in connection with the rocks of the Berlin valley; even in these rocks, the mineral is rarely prominent. The greatest concentration is shown in a thin lens, about 10 feet east of the thrust fault that terminates the limestone mass, southeast of Center Berlin. The greenish-gray, nearly aphanitic rock is penetrated by what appears to be veins, but under the microscope a finely scaly groundmass of quartz and sericite is seen mixed with a mushy, dark-grayish substance that may be partly graphite, partly chlorite. Under crossed nicols, the rock displays hundreds of equidimensional albite crystalloblasts, about .1 mm. across, that enclose the schistose groundmass, but most of their margins have remained clear, and the groundmass is absent from them (Pl. 11, figs. 1, 2). Likewise, what appear to be veins, megascopically, are bleached zones, from which only the dark pigment has been removed, but the sericite-quartz groundmass continues through them in unchanged direction. A few small albite crystalloblasts were noted in quartz-chlorite schist from summit 1820′, 1.1 miles northeast of Center Berlin, but elsewhere in this very large mass of schist the mineral seems to be practically absent. Twelve specimens were examined from the cuts of

PLATE 8.—PHOTOMICROGRAPHS OF ROCKS FROM RENSSELAER GRAYWACKE AREA

FIGURE 1.—COARSE FELDSPATHIC QUARTZITE, INTERBEDDED WITH PURPLE AND GREEN SLATE

About 100 stratigraphic feet below graywacke layers, 1 mile northwest of Davitt Pond, Troy quadrangle, New York. Large clastic quartz grains; a = albite, c = chloritized biotite. Ground-mass largely chlorite, sericite, and small quartz grains. Plane-polarized light.

FIGURE 2.—TUFFACEOUS LENS IN GREENISH-GRAY SILICEOUS SLATE

About 40 feet below basal graywacke layers at hill 1454′, 1¼ miles west of Petersburg, New York. Rhombs of calcite replace fine-grained groundmass of chlorite, quartz, albite, and enclose magnetite crystals. Plane-polarized light.

FIGURE 3.—BASALT PEBBLE IN CONGLOMERATE LENS

Hill 1240′, ¾ mile southeast of Red Rock settlement, Kinderhook quadrangle, New York. Prominent laths of albite (ab_{96}), surrounded by fine-grained aggregate of chlorite, secondary magnetite, sericite, and carbonate replacing pyroxenes. Some dark grains are ilmenite with shells of sphene. Plane-polarized light.

FIGURE 4.—VARIOLES IN BASALT PILLOW

Half a mile west of Banker Pond, Grafton quadrangle, New York. Radiating needles of albite, or a zeolite, are crossed by veins of epidote, quartz, chlorite, and albite in various combinations. Plane-polarized light.

FIGURE 5.—ALBITE BASALT DIKE IN GRAYWACKE

At Turner Mountain, Stephentown Center quadrangle, New York. Large white grain in lower right area is pyroxene, surrounded and partly replaced by chlorite, sericite, epidote, secondary magnetite, and quartz. Smaller white grains are remnants of albite. Several large black grains are ilmenite and sphene. Plane-polarized light.

FIGURE 6.—GRAYWACKE

Showing epidote needles penetrating quartz (light grain in center) and albite (top part of light grain) near quartz-epidote vein. Dark shell surrounding quartz is fine-grained aggregate of epidote and chlorite. Snake Hill, ¼ mile northwest of Barberville, Troy quadrangle, New York. Plane-polarized light.

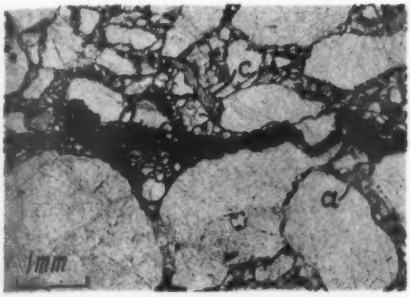


FIGURE 1

FIGURE 2

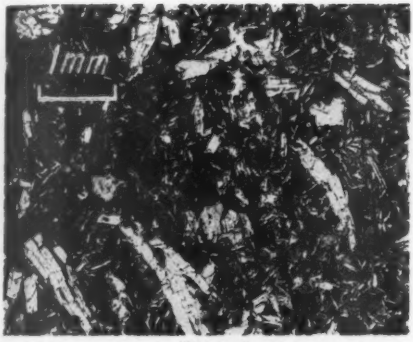


FIGURE 3

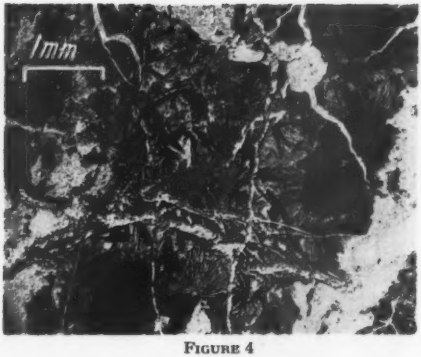


FIGURE 4

FIGURE 5

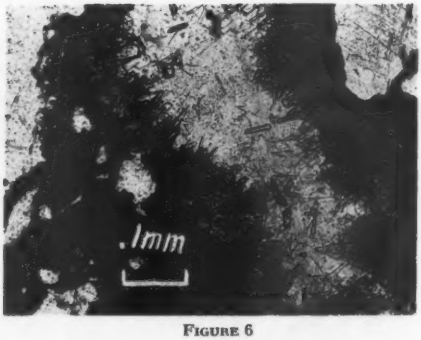


FIGURE 6

PHOTOMICROGRAPHS OF ROCKS FROM RENSSELAER GRAYWACKE AREA

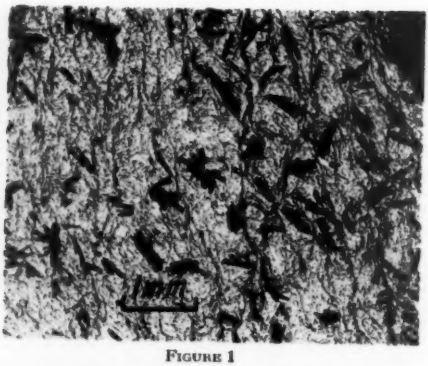
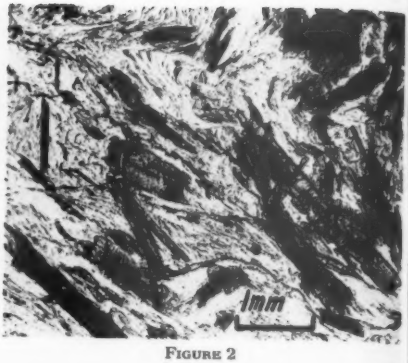


FIGURE 1



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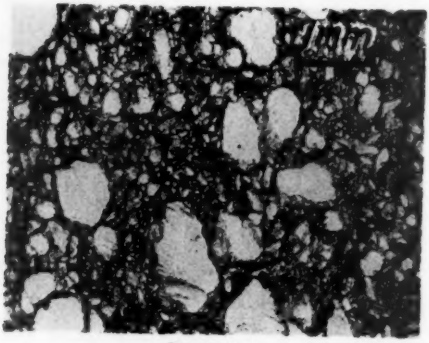

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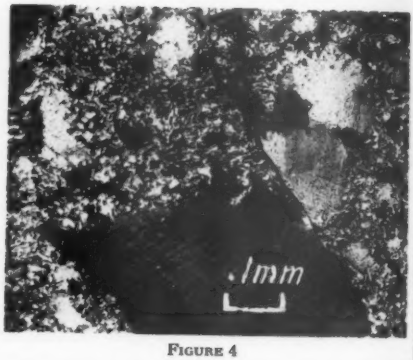

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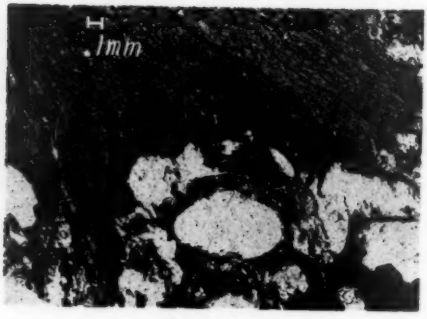

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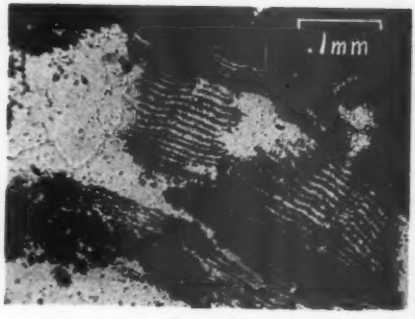

FIGURE 6

PHOTOMICROGRAPHS OF SCHIST FROM TACONIC RANGE

State highway No. 2, on the northeast slopes of hill 1719′, ½ mile west of Bee Hill (7½′ Williamstown quadrangle, Mass.-Vt.). These outcrops are only 1¼ miles west of the east base of the Taconic Range, and yet albite is conspicuously absent from nearly all slides. Figure 4 of Plate 11 shows the most albitic rock found.

Epidote, so common in the graywacke areas, is nearly absent from the Taconic Range. One exception is a tract of highly epidotized tuff, 1½ miles west-northwest of Hancock. The southernmost outcrops (where a power transmission line crosses ridge 1240′) are within 200 feet of the thrust fault east of the limestone tract of the southern Berlin valley and show extraordinarily strong deformation of the rock into a quartz-chlorite-epidote mylonite. Lamination and *a* lineation plunge at about 30° E., and some sections of the rock are a solid lemon green and consist of myriads of minute epidote grains, in thousands of parallel streaks, intermixed with quartz. Surrounding these epidotic layers are dark-green ones of chlorite, quartz, some albite, and quantities of magnetite, mostly in octahedra distorted into little plates parallel to the direction of the lineation. White films and wider lenses between the chlorite-magnetite rocks are largely quartz and albite (ab_{98}; Pl. 11, fig. 5). Boulders of a fine-grained, gray quartzose rock, as much as a foot across, are found sporadically in this mylonite. They too are drawn out into spindles parallel to the *a* lineation and may represent pyroclastic material.

Structure.—Exposures in the Taconic Range are generally poor. The ledges are low, discontinuous, and lack depth from which the continuations of individual beds could be gleaned. However, the valley of Hoosic River affords a fairly good cross section, at least on its north bank, and, 5 miles south, a newly blasted 4-mile stretch across the crest of the range, where State Highway 2 crosses it, gives a very satisfactory picture of the structure of the range. Both cross sections show substantially the same: an endless sequence of small, open, westward-overturned folds, with axial-plane cleavage dipping east, or southeast, at angles of about 60°, and innumerable slip planes and shear zones, dissecting the entire range (Pl. 1, cross sections 4–9). That the motion along these slip planes has been obliquely up from the east-southeast can be shown where recognizable beds, or quartz veins, have been displaced. Drag along the planes is relatively rare. In places, the main shear zone shows none, but subsidiary fractures with small displacements may show it. Some of these shear and slip zones may be normal faults.

Despite considerable attention to detail, no recumbent folds have been found anywhere in the range. Even isoclinal folds are rare. If the shear zones are considered absent for a moment, the range would resemble a huge corrugated sheet which dips at about 20°–30° E. Unfortunately, there seem to be no marker beds in the argillaceous series, so it is impossible to say to what extent this corrugated sheet has been

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PLATE 9.—PHOTOMICROGRAPHS OF SCHIST FROM TACONIC RANGE

FIGURE 1.—TYPICAL QUARTZ-CHLORITE SCHIST OF THE TACONIC RANGE

Small crystals of chloritoid, either singly or in radiating groups, embedded in fine-grained groundmass of chlorite, quartz, and sericite. *s* planes are marked by thin films of sericite (top to bottom in photograph). Summit of Bald Mountain, 3½ miles northeast of Petersburg, New York. Plane-polarized light.

FIGURE 2.—COARSE QUARTZ-CHLORITE SCHIST WITH LARGE CRYSTALS OF CHLORITOID

Enclosing in places tracts of graphitic dust of the groundmass as internal *s*. Summit of hill 1540′, ¾ mile north of Mason Hill, Bennington quadrangle, Vermont. Plane-polarized light.

FIGURE 3.—GRAYWACKE FROM 8-INCH LAYER IN GREENISH QUARTZ-CHLORITE SCHIST

North spur of rise 2135′, 1½ miles northeast of Center Berlin, New York. Grains of quartz surrounded by felt of sericite, chlorite, and quartz. Boundaries of several quartz grains indistinct because of incipient penetration by sericite scales growing out from recrystallized groundmass. Plane-polarized light.

FIGURE 4.—SAME ROCK AS IN FIGURE 3

Showing sericite in single scales and veinlets penetrating microcline grain (below center) and quartz (above and to right of center). Crossed nicols.

FIGURE 5.—CONGLOMERATIC GRAYWACKE IN QUARTZ-CHLORITE SCHIST

From ravine on west slope of rise 2458′, 3¼ miles north-northeast of Stephentown, New York. Grains of quartz are embedded in groundmass of chlorite and graphite. Note minute films of graphite penetrating quartz grain in center. Plane-polarized light. *See* Figure 6.

FIGURE 6.—ENLARGED VIEW OF CENTRAL AREA OF FIGURE 5

To show graphite films in quartz. In each quartz grain, the films follow a uniform direction.

thickened and laterally compressed by the many shear zones. If there were a simple correlation between the thickness of a sheared zone and the amount of movement on the zone, one might attempt an estimate, but there is no reliable index. The most intense deformation of the rocks, as seen in the very thoroughgoing mashing, *a* lineation, and recrystallization as described from the epidote-quartz-chlorite mylonite west of Hancock, is absent within the range, in fact so much so that it requires a little training to see the *a* lineation along some of the shear zones. The only isograd in the range, that of chloritoid, shows no sudden offset, but, as it strikes parallel to the strike of the range, this feature, too, is of no avail.

Though our knowledge of the internal structure of the Taconic Range is limited, we may examine its structural relations to the surrounding areas; beginning in the north, we first discuss the zone from Bennington to North Petersburg.

AREA MOUNT ANTHONY—NORTH PETERSBURG:

The northeastward-trending limestone belt of Breese Hollow has been mentioned. East of its northeast end stands Mount Anthony, underlain by quartz-chlorite schist, and the entire southeast border of the limestone tract is accompanied by this schist. Walcott and Prindle have studied Mount Anthony; the mountain is, essentially, a southward-plunging syncline. The Breese Hollow limestone tract dips under it from the northwest, and Walcott (1888, p. 237, 238) found Beekmantown fossils on both the northwest and east sides of the mountain.

Plate 1 shows two limestone tracts on the northwest slope of Mount Anthony. On the north side of the mountain, a black, rather siliceous slate separates them, but father southwest greenish slate appears between the two limestone zones. The southeastern limestone seems to end just west of the New York-Vermont State line. However, so much till covers the forested area to the west that the limestone might continue as a much thinned plate, or a series of small lenses, without being noticed in the field. At any rate, 1½ miles southwestward along the line of strike, limestone reappears as a plate, about 20 feet thick, dipping southeastward, and by means of several other outcrops this limestone band could be extended southwestward across Hoosic River.

Locally both the limestone of this southeastern belt and the adjoining quartz-chlorite schist of the Taconic Range show the effect of appreciable deformation, and the east-southeast-plunging *a* lineation remains one of the most reliable criteria. Yet the outcrops with this deformation structure are not so simply distributed that a single, smooth line could be drawn to indicate a thrust. Instead, the evidence favors the idea of a fairly wide zone of upthrusts. In one place, the principal dislocation may lie between the quartz-chlorite schist and the southeastern limestone. In another place, the deformation of the argillaceous rocks between the limestones is more intense. At still other places, the limestone is isoclinally folded, strongly lineated whereas the nearest outcrops of slate or schist are relatively free from strain. Though there can be no doubt that thrusting has occurred along this curving zone, its amount remains largely unknown. This dilemma is probably characteristic for the entire vicinity: outcrops are small and widely scattered, and most of the rocks are so incompetent and devoid of any key beds that the true magnitude of the displacements cannot be ascertained. A few details may show the nature of the problems.

On the north slope of Mount Anthony, lineation plunging east-southeast is strongest at the top of the southeastern limestone, apparently absent in the slate below it, and sporadically developed in the lower limestone, although, as noted above, this rock is fossiliferous. Between the Bennington and North Pownal quadrangle boundaries and the New York-Vermont State line, both limestones display intense schistosity with *a* lineation along narrow zones in which the layers of deposition have been isoclinally folded. Even recumbent position of the limbs was noted, for instance at the little spur, 1020′, 1 mile southwest of the northeast corner of North Pownal quadrangle, and in cliffs of the southeastern limestone, about a mile south-southeast of the spur just mentioned. But between these are others in which the details of bedding, with thin sandy layers, are well preserved, and little if anything suggests intense deformation near by.

Loc. 25. At its western termination, just west of the State line (hill 1580′, 1.6 miles east-northeast of School No. 11, North Pownal quadrangle, N.Y.-Vt.), the southeastern limestone is severely deformed. Bedding has for the most part been obliterated by slip planes along which the rock has been drawn out in a west-northwesterly direction, and intense *a* lineation accompanies it. The crushed rock has been recemented and in part silicified, and it appears that some of the normally overlying quartz-chlorite schist has been kneaded into the base of the limestone in a manner that recalls the relations at Butternut Hill (Pl. 7, cross section 3). About 300

feet west of where the limestone disappears, the black slate, which so commonly accompanies the Ordovician limestone, is intermixed with the green quartz-chlorite schist, presumably in the exact prolongation of a thrust zone. Yet the two argillaceous rocks show no unusual strain. Layers of deposition are recognizable, especially in the green schist, where they show as gently corrugated, nearly horizontal planes, dipping south.

Compared with these outcrops, the limestone 1½ miles southwest is nearly undisturbed. Bedding is well recognizable, tilted to the southeast, but neither folded nor lineated, nor sheared through by slip zones. Even the surrounding chlorite schist with graywacke lenses shows very little deformation.

The exact relationship between the two limestone belts southeast of Breese Hollow is obscure. On the northwest slope of Mount Anthony, one gains the impression that this is a primary sequence, with slate in between. But in the portion of the belt within New York State, the intervening argillaceous rock is the green and purple quartz-chlorite schist, with characteristic graywacke lenses (well exposed, for instance, at the little knoll 1300′, 1 1/10 miles east of Indian Hill). One could therefore assume a thrust fault above (southeast of) the main limestone tract, along which the quartz-chlorite schist has overridden the limestone series. Somewhere to the east, this thrust should presumably connect with the thrust southeast of the southeastern, perhaps stratigraphically higher, limestone. But in view of the poor exposures, these are all speculative interpretations.

AREA NORTH PETERSBURG—HANCOCK: The thrust zone just discussed seems to continue southward along the Berlin valley, but north of Berlin the exact location and magnitude of displacement along the zone are so difficult to estimate that we have not marked the hypothetic trace of the thrust on the map.

One mile south of North Petersburg, the various limestone belts of the Breese Hollow and Indian Hill section end, but overburden obscures the structural relations. The western limestone outcrop, on State Highway 22, is probably terminated on the west by a high-angle normal fault, but the eastern outcrop (1.2 miles southeast of North Petersburg, on the northwest slope of the northern summit, 1360′, North Pownal quadrangle, N.Y.-Vt.) may lie on a thrust. The small outcrop shows massive, joint-scarred limestone, about 100 feet thick, but along the lower, western side a few small ledges exhibit eastward-dipping *s* planes and weak *a* lineation. Small ledges of gray, silty slate with graywacke lenses, east and west of the limestone, show no traces of conspicuous deformation.

The thick zone of siliceous schists and quartzite lenses that passes through Odell, Moon, and Poplar Hill, next to the south, shows east-southeast-plunging lineation and even a few axes of folds in that direction. If a thrust zone passes this area it is likely to lie east of the arenaceous zone because the latter is obviously continuous with the siliceous green and purple slate belt that crosses Little Hoosic River. Nearly everywhere the chloritoid isograd lies about ¼ mile east of the trace of this hypothetical thrust zone and might therefore be genetically connected with it, particularly in view of what was said above about the age relations of this mineral. Large loose blocks of dolomite and limestone on the east-west trending crest of Odell Hill, ⅛ mile east of the summit, may possibly mark another remnant of the limestone zone southeast of North Petersburg. If in place, the rock would help further to mark the trace of the fault zone. On Plate 1, it has been shown with a question mark.

In the eastern slopes of the valley, beyond the siliceous rocks just mentioned, the greenish-gray colors continue in the quartz-chlorite schists; quartzitic and graywacke lenses are occasionally seen, on both sides of the chloritoid isograd, and no particular line or zone marks a definite lithologic break. Nevertheless, as the structure symbols on Plate 1 show, thrusting and stretching of the rocks in the familiar direction, obliquely up to the west-northwest, is very generally recorded, both on tilted and folded bedding planes and on slip cleavage planes. The only way to reconcile this situation, it seems, is to assume a zone of distributive thrusting, perhaps half a mile wide, with displacements sufficiently small to leave the lithologic sequence of rocks essentially uninterrupted. If the westernmost quartzose slates, with the thin limestone interbeds, near Moon Hill, are Lower Cambrian, it could be assumed that the bulk of the rocks of the range are also of that age, in agreement with Prindle's interpretation (Prindle and Knopf, 1932, p. 290).

East and south of Berlin, the trace of the thrust zone is well marked by the juxtaposition of the green quartz-chlorite schist on the one side, and black slate and limestone on the other. The area, 1–3 miles north-northeast of Berlin, was not surveyed in detail, and if there are outcrops of black slate and limestone the fault can be extended several miles in that direction. The throw along this thrust fault may be con-

siderable, or it may be moderate if the interpretation of the Butternut Hill thrusts is approximately correct, where the green schist may be present at rather shallow depth beneath the black slate and limestone. Also, if the tuff zone west-northwest of Hancock, and the albitized tuff southeast of Center Berlin, were originally continuous with the similar tuff lenses on the west side of the Berlin valley, one might argue that the thrusting has not separated these parts excessively; but exact figures cannot be given.

About a mile southeast of Hancock, Massachusetts (Pl. 1), a northeast-trending thrust fault zone lies east of the limestone tract that underlies the valley northeast of the village. The extremely deformed limestones are accompanied on the east by siliceous greenish slates that contain lenses of graywacke and a magnetite-bearing chlorite schist that resembles the outcrops west of Hancock so closely that we regard it as another tuffaceous lens. Good exposures showing the intensely deformed condition of the rocks are at the southwest base of Potter Mountain (7½′ Hancock quadrangle, Mass.-N.Y.). Probably this thrust zone joins the one south of the Berlin valley, but the critical area has not been examined in detail.

We turn now to the rocks east of the Taconic Range.

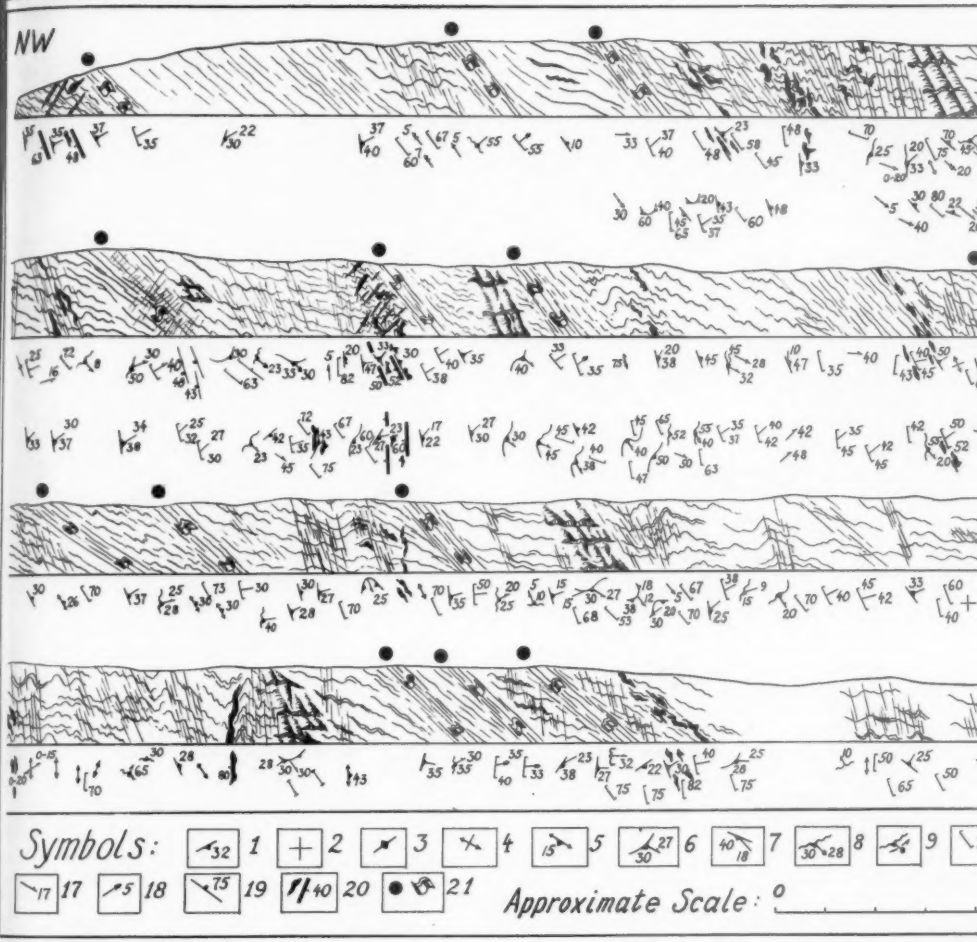
4. North Pownal Valley

The eastern base of the Taconic Range is underlain by a belt of limestone and black slate at least in part, and perhaps entirely, of Ordovician age. It is marked by a valley, but north of North Pownal, Vermont (Pl. 1), Hoosic River breaks northwestward across the Taconic Range whereas the limestone continues northward for about 2 miles, and then ends with two parallel branches. The western contact of the limestone is a thrust fault along which the limestone has overridden the flooring quartz-chlorite schist. This thrust is excellently exposed just west of a large cement quarry at North Pownal, and a long railroad cut, south of North Pownal station, as well as smaller ones along State route 346, give insight into the structure of the schist a few hundred feet below the thrust.

The thrust contact is exposed intermittently on the west slopes of hills 1140′, 900′, and 860′, directly southeast of the village of North Pownal. On the east side is a bold cliff of limestone, locally folded and kneaded into the gray or black, finely fissile slate so commonly associated with the Ordovician carbonate rocks of this area. On the west, flooring the latter, are ledges of the green chlorite-quartz schist, commonly with microscopic chloritoid needles. As a whole, the thrust plane dips about 30° E.-NE., but deviations, both in strike and dip angle, exceed 35°. Over distances of 3–10 feet, the thrust is a smooth plane, but on a larger scale it is warped, variously flexed, and, it seems, offset here and there by later high-angle faults of east-west strike. The overlying limestone is rarely massive, commonly folded isoclinally, but the orientation of the axes of small folds is variable, and in favorable places one can see that folds of different order of magnitude are co-ordinated and superposed on each other, so that a large mass of small folds may wrap around the flank of a large fold or may fill the irregular space between fractured (presumably dolomitic) portions of the rock. Commonly the limbs of small folds are sheared through, displaced, and thoroughly recrystallized (Fig. 17). Even some of these healed ruptures have subsequently been deformed. Small-scale rupturing and faulting has distorted dolomitic blocks, whereas the surrounding calcitic limestone may show nothing of the kind (Fig. 18), even though its total deformation is probably much more intense.

In the underlying quartz-chlorite schist a very different style of deformation is apparent. Even 2 feet below the separation plane, in places thin quartzose layers, only slightly contorted, show open folds a little overturned to the west, and crossed by axial-plane cleavage dipping at 80° E., the whole crossed by chlorite seams and irregularly shaped quartz veins. This picture of trifling deformation, however, changes within 5 or 10 feet upward. The layers of deposition become obscure, are thinned out, the fragments show violent shearing combined with tight isoclinal folding, and may be completely lost in a mass of smeared-out shreds of mica and quartz. The entire rock is pervaded by the east-southeast-trending lineation which is equally prominent in the schist and the overlying carbonate rocks. Even larger masses of the schist have accommodated themselves to this direction of strain and appear as limbs of folds, several feet long, in such orientations that the axes of these folds plunge east-southeast parallel to the *a* lineation. Thus, a large limb on the west slope of hill 1140′, directly below the thrust, strikes east-southeast in perpendicular position.

In places, blocks of dolomite over 4 feet across have been squeezed into the chlorite schist which, in turn, has been smeared into lenses of the black slate so that for several inches one may be in doubt where to draw a boundary. A good exposure, showing also the strongly undulating surface of the thrust, projects through the west base of the talus of quarry debris, about 140 feet east of the highway. The chlorite schist, directly below the thrust, is also well exposed in a small blasted ledge, about half way between the road leading from the highway to the quarry and the highway cut south of the quarry. The ledge is only 30 feet east of the highway, but hard to see because of underbrush.

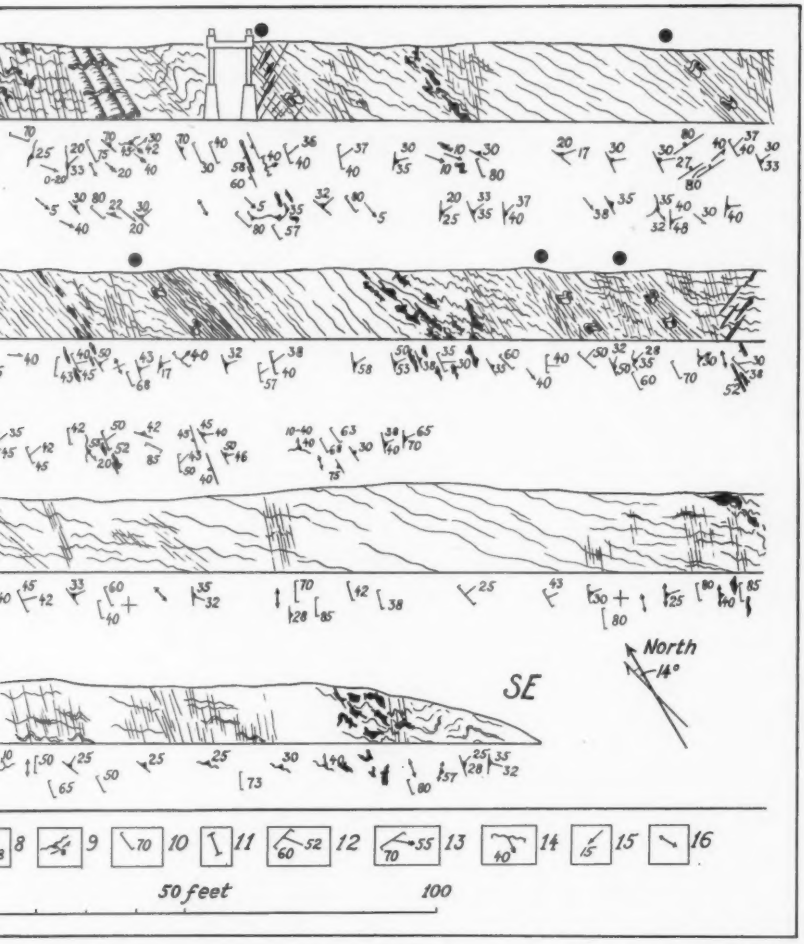


STRUCTURE OF QUARTZ-CHLORITE SCHIST, IN RAILROAD CUT SOU

Continuous profile shown in four strips, accompanied by stru

1. Strike and dip of beds
2. Horizontal beds
3. Vertical beds
4. Horizontal beds, arrow shows bearing of axes of folds
5. Beds dip 15°, axes of folds horizontal
6. Curved beds, with bearing and plunge of axis of fold
7. Strike and dip of beds with trend and plunge angle of lineation
8. Contorted beds, with bearing and plunge of both axes of folds and lineation
9. As in 8, older folds and lineation crossed by later lineation
10. Strike and dip of slip cleavage, or of *s* planes of unknown origin
11. Vertical slip cleavage or *s* planes
12. Strike and dip of slip cleavage or *s* planes, with bearing and plunge of lineation
13. Slip cleavage or *s*
14. Contorted *s* planes
15. Axis of fold, bear
16. Horizontal axis o
17. Bearing and plun
18. Bearing and plung
19. Strike and dip of
20. Quartz in irregula
21. Black dots denote with axes of small fol zone.

Folds plunging st



CUT SOUTH OF NORTH POWNAL STATION, VERMONT

mpanied by structure measurements

lip cleavage or *s* planes, with bearing and plunge of axes of folds and lineation
ontorted *s* planes, with bearing and plunge of axes of folds and lineation
xis of fold, bearing and plunge angle
Horizontal axis of fold
Bearing and plunge angle of lineation
Bearing and plunge angle of axes of folds and lineation
Strike and dip of joint
Quartz in irregular masses, or veins, with dip angle
Black dots denote subsidiary thrust zone, with *a* lineation trending east-southeast and, commonly, axes of small folds having same orientation, as shown diagrammatically by second symbol in slip

Folds plunging steeper than 25° SW. shown by diagrammatic sketch.

A cut of the Boston and Maine railroad, about 1200 feet long, starts south of North Pownal station and gives one of the most instructive views of the structure of the quartz-chlorite schist, several hundred stratigraphic feet below the thrust just described. The cut runs southeast-northwest so that special symbol in the shear zones showing their orientation.

Commonly, west-dipping cross joints, at right angles to the *a* lineation and fold axes, penetrate the sheared zones; quartz veins may follow them and bring about local silicification of the sheared rocks.

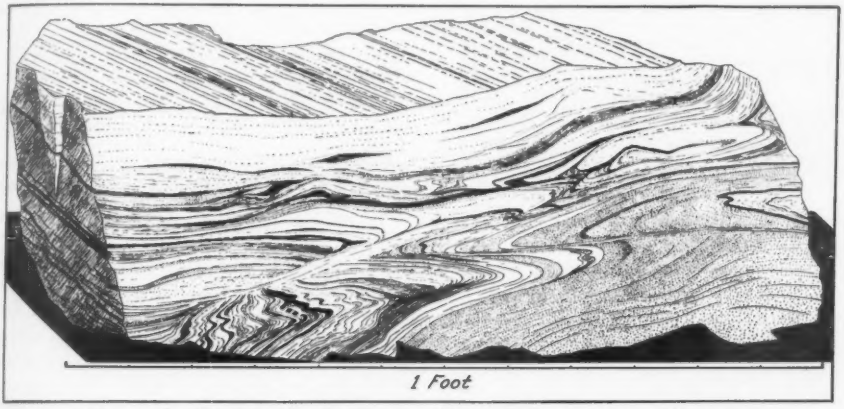


FIGURE 17.—GRAPHITIC LIMESTONE FROM QUARRY, NORTH POWNAL, VERMONT

Showing folding, shearing, and recrystallization of calcitic zones. Axes of folds are parallel to lineation which shows on top surface.

the southeast end is somewhat closer to the thrust than the northwest end. The writer has mapped the structure and believes that it illustrates reliably the behavior of a series of incompetent beds subjected to thrusting (Pl. 10).

One may distinguish in the schist a "regional" and a "thrust" structure. The regional structure consists of innumerable small folds, like the hundreds that are exposed along State Highway 2 on both sides of Petersburg Pass. Their axes strike north-northeast, the limbs are mildly overturned to the west-northwest, and wave lengths rarely exceed 5 or 6 feet (Pl. 6, fig. 2). The axes of the folds plunge at low angles in either direction. Local plunges of as much as 30°–40° are noted as uncommon and have been shown on Plate 10 by a special diagrammatic sketch in the section (one place in each of the four strips). Axial-plane slip cleavage, not everywhere present, dips at high angles east to east-southeast and produces the customary north-northeast trending *b* lineation where it crosses the limbs of folds.

At 17 places, marked by black dots, this regional structure is interrupted by slip zones, 1 to 30 feet wide, inclined to the east-southeast. In each of them, the ordinary structure elements are superseded by intense schistosity dipping about 30°–50° E. or E.-SE., and *a* lineation plunging very nearly down-dip. The original layers of deposition of the schist are rarely visible within these slip zones. Where they are, one may recognize them amid the mass of slivers and slabs of deformed schist as delicate rods or twisted lenses, and, wherever found in place, the axes of their corrugations plunge east-southeast, parallel to the *a* lineation (Pl. 6, fig. 3; Pl. 11, fig. 6). To emphasize this deviation in orientation of the fold axes, we have inserted a

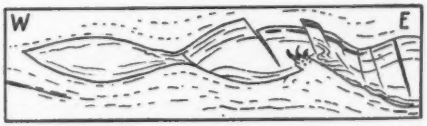


FIGURE 18.—CALCITIC LIMESTONE, 1 FOOT ABOVE THRUST FAULT AT WEST BORDER OF NORTH POWNAL LIMESTONE BLOCK

Showing fracturing and displacements of lens of dolomitic limestone. Length of surface, 1 foot.

The foregoing description is very brief. Any one who examines this interesting railroad cut in detail will discover that there is more than one cleavage system; that the discrimination among bedding, slip cleavage, fracture cleavage, and other slip planes is difficult; that quartz veins of many ages can be differentiated; and that the separation of "regional" and "thrust" structure is by no means as simple as shown in the cross section. He will find many other faults with the present representation. But undoubtedly he will verify the fact that the schist mass is dissected into numerous oblique plates, or huge lenses, by eastward-dipping slip zones. If the observations are approximately correct, there are at least 17 of them within something like 600 stratigraphic feet of the schist. If this is the normal number (and the profile across the Taconic Range along Highway 2 shows a fairly similar spacing of slip zones), it may mean that the entire schist complex is penetrated by hundreds of them. Between them, the layers of deposition still show innumerable small-scale contortions, and the microscope reveals addi-

tional hundreds of thousands. How far individual blocks have slipped past each other remains unknown in the absence of key beds, but presumably thrust zones in this schistose rock are very wide indeed, and the deformation has been imparted to impressive volumes of rock. Reduced in scale, the picture of the deformation would resemble the nearly homogeneous strain of a soft, ductile, and thoroughly weak substance yielding in one direction. For a comparison of the deformation on various scales, it is instructive to compare Plate 10 with the photomicrograph (Pl. 12, fig. 3).

The schist exposures near the thrust zone at North Pownal seem to justify the belief that thrust zones in this rock are reliably recorded by a number of elements of deformation. Although an individual ledge may show practically no deformation, exposures scattered over larger areas disclose a systematic arrangement of a number of characteristic features. If they are mapped over a larger territory, they should enable us to trace the zones of maximum deformation with considerable accuracy. This is a result of general value; it means that thrust zones can be recognized, even where exposures are poor and few. Therefore, it seems improbable to the writer that in this region a thrust of the first order of magnitude could remain unnoticed.

Little can be said about the floor upon which the limestone was deposited. However, in the northern prolongation of the western limestone branch (in the bed of Potter Hollow, 2 miles north of North Pownal, locality 26) a series of southeast-dipping black slates and dark-gray quartz schists is interbedded with at least six lenses of hard, graphitic, and quartzose limestone (Pl. 1). The lenses, not over 2–3 feet thick, are scattered through about 200 feet of the section. The quartz schists contain no chloritoid and little chlorite, but, curiously, little calcite veinlets that traverse the impure limestone lenses contain crystalloblasts of nearly euhedral albite (ab_{92}), 1–2 mm. in size. These rocks appear to rest on the quartz-chlorite schist that underlies the northernmost part of the Taconic Range, but lithologically they differ from them. Quartz-chlorite schist is also exposed ¾ mile south of these exposures where the thrust zone, just described, is lost under varved clays in the south-flowing brook to which Potter Hollow Brook is tributary. The possibility should be considered that these strata are a series of basal Ordovician beds, since separated from the main mass of limestone and black slate by the thrust zone of North Pownal.

PLATE 11.—PHOTOMICROGRAPHS OF METAMORPHOSED ROCKS OF TACONIC RANGE

FIGURE 1.—LENS OF FINE-GRAINED METATUFF IN QUARTZ-CHLORITE SCHIST

One mile east of Center Berlin, New York. Finely laminated groundmass of sericite, quartz, and chlorite, with films of graphite, enclosed by equant albite crystalloblasts whose clear margins show as discontinuous white streaks. Plane-polarized light. Compare Figure 2.

FIGURE 2.—SAME AREA AS SHOWN IN FIGURE 1

With crossed nicols to show pattern of albite crystalloblasts.

FIGURE 3.—CONCRETIONARY MASS OF Fe-Mn CARBONATES IN QUARTZ-CHLORITE SCHIST

West slope of Berlin Mountain, 7½′ Berlin quadrangle, N. Y.-Mass.-Vt. Groundmass of chlorite, graphite, sericite, carbonate, and quartz is replaced by large rhombohedra of ankerite.

FIGURE 4.—BLACK PHYLLITIC SLATE THAT OVERLIES QUARTZ-CHLORITE SCHIST

On State Highway 2, eastern border of 7½′ Berlin quadrangle, N. Y.-Mass.-Vt. Bedding, gently folded, slopes to left. Quartzose layers at bottom. In top portion, small albite crystals show as white and black dots between fine-grained groundmass of sericitic muscovite, quartz, and some chlorite. Crossed nicols.

FIGURE 5.—QUARTZ-ALBITE LENS IN MYLONITIZED EPIDOTE SCHIST

Probably of tuffaceous origin. Quartz in minute granulated, partly recrystallized grains surrounds larger grains of twinned albite showing minute faulting and bending. Hill 1240′, 1¼ miles west-northwest of Hancock, Mass.

FIGURE 6.—QUARTZ-CHLORITE SCHIST OF ROCK SHOWN IN FIGURE 3 OF PLATE 6

Section across *a* lineation shows axes of microscopic folds parallel to *a*. Note chloritoid crystals, to right of center, apparently fractured and displaced by deformation of schist. Other, small fractured crystals in dark chloritic beds near center of area. Plane-polarized light.

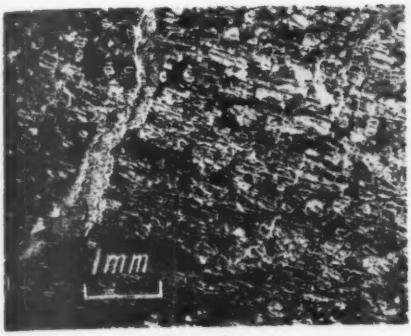


FIGURE 1

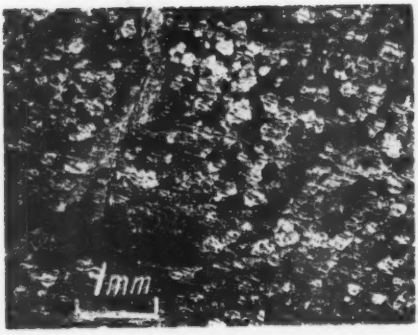


FIGURE 2

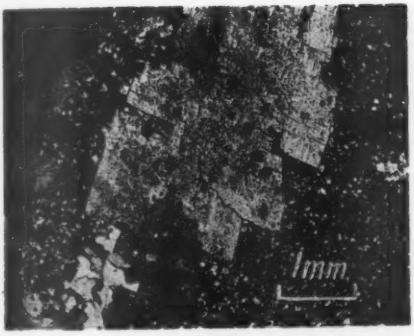


FIGURE 3

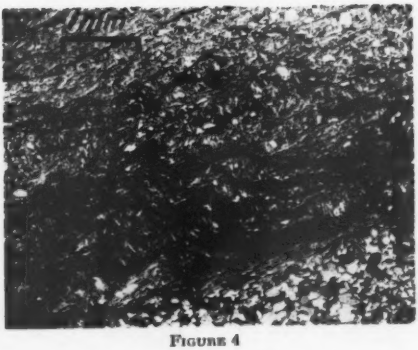


FIGURE 4

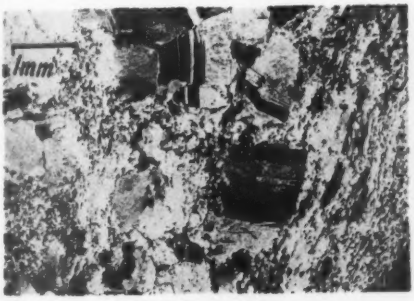


FIGURE 5

FIGURE 6

PHOTOMICROGRAPHS OF METAMORPHOSED ROCKS OF TACONIC RANGE

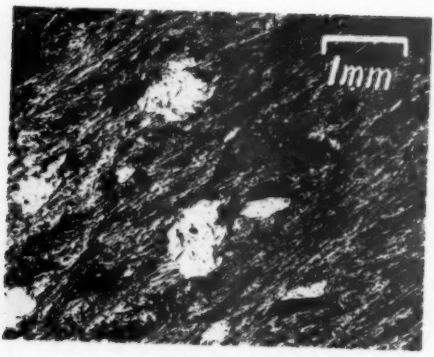


FIGURE 1

FIGURE 2

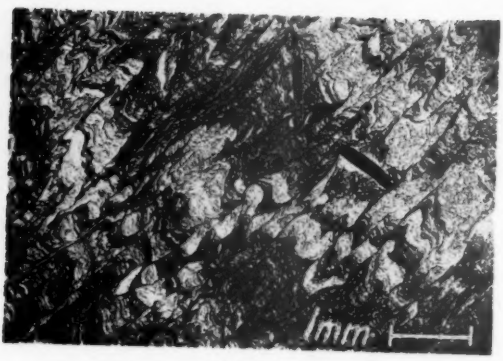


FIGURE 3

PHOTOMICROGRAPHS OF METAMORPHOSED AND DEFORMED SCHIST

As Plate 1 shows, the quartz-chlorite schist of the Taconic Range separates the North Pownal limestone zone from the Breese Hollow and other northern carbonate belts. Whether the Ordovician rocks have been cut out on the north by a fault zone, or whether they were perhaps not deposited north of the Potter Hollow exposures is unknown. An isolated group of limestone boulders, (shown with a question mark on Pl. 1) in the northernmost section of the brook that drains the north-south valley north of North Pownal may or may not indicate an additional outcrop, hidden under Quaternary cover.

Valley deposits and till conceal the southern continuation of the thrust zone of North Pownal. The writer has not examined in detail the east slopes of the Taconic Range southwest of North Pownal, and whether Ordovician sediments were ever deposited west of the thrust zone is unknown. However, the last ¾ mile of cuts along State Highway 2, east of the crest of the range, is in black phyllitic slates with minute albites (Pl. 11, fig. 4), and their superposition on the green quartz-chlorite schist is exposed (200 feet northwest of the road intersection, at elevation 1564′, ¼ mile west of the eastern border of 7½′ Berlin quadrangle). At the contact, the black slate contains lenses of a dark-gray, impure limestone. The rocks do not resemble any types seen in the schist complex, and one wonders whether this group of rocks may be a thin plate of Ordovician rocks resting on an older quartz-chlorite schist, and perhaps once connected with the limestone-slate series in the North Pownal valley.

The eastern contact of the limestone belt of North Pownal is concealed, but we shall refer to it again in connection with the next area.

5. *Pownal Upland*

General description.—A tract of high, level ground that extends southward for 7 miles from Mount Anthony, to connect with the Green Mountain front, may be called the Pownal upland, in reference to the township of that name in southwesternmost Vermont. Scenically and geologically, it is a unique area for it bridges the two ranges, and from its elevated edges, magnificent views are had, northward across the termination of the Rutland-Bennington limestone valley of Vermont, to the sweeping skyline of the northern Taconic Range, culminating in Mount Equinox (Pl. 2, fig. 1); farther west and southwest over the southern Taconic Range (Pl. 2, fig. 2); and southward to the heights of Mount Greylock. Only on the east, the somber forests of the Green Mountain front surmount the upland and close the view.

The upland is underlain by a flat plate of quartz-chlorite schist; it might be called a plateau were it not for slight undulations and flexures of the bedrock, so that the surface corresponds closely, but does not exactly agree, with the bedding of the formations.

Lithologic characteristics.—Probably the most conspicuous rock on the upland is a greenish schist, much more quartzose than the ordinary quartz-chlorite schist of the Taconic Range, and in places the rock is a chloritic quartz schist (for instance, along the crest of a ridge, ½ mile southwest of the summit of Carpenter Hill, and on the east-facing scarp, just north of Pownal Center, Bennington quadrangle).

PLATE 12.—PHOTOMICROGRAPHS OF METAMORPHOSED AND DEFORMED SCHIST

FIGURE 1.—QUARTZ-CHLORITE SCHIST WITH ALBITE PORPHYROBLASTS

From north slope of Mann Hill, Bennington quadrangle, Vermont. Twinned and untwinned albite embedded in finely scaly groundmass of chlorite, sericitic muscovite, quartz, and magnetite. Crossed nicols.

FIGURE 2.—QUARTZ-CHLORITE SCHIST

At limestone contact shown in Figure 4 of Plate 6. Groundmass of recrystallized muscovite, chlorite, and quartz, holding porphyroblasts of albite. Crossed nicols.

FIGURE 3.—QUARTZOSE PHASE OF QUARTZ-CHLORITE SCHIST

Half a mile southwest of summit of Carpenter Hill, Bennington quadrangle, Vermont, showing small-scale slippage and shear folding of beds. Thin plates of magnetite or ilmenite nearly parallel to bedding. Plane-polarized light.

Greenish, slaty phases of the chlorite-quartz schist resemble the chloritoid-bearing quartz-chlorite schist of the Taconic Range, but, with the exception of two specimens from hill 1540′, 1 mile southeast of Mann Hill (Bennington quadrangle), none of the specimens examined contain chloritoid. However, the argillaceous rocks of the upland contain small biotites (Pl. 1) and, in the southernmost area, albite crystalloblasts (Pl. 12, fig. 1). The two isograds have been drawn on the basis of about 80 thin sections from this part of the upland. Magnetite crystalloblasts, up to 7 mm. in diameter, are abundant on the north slope of Mann Hill, and on hill 1540′, 1 mile southeast of it. Black, graphitic quartz schists, so fine-grained that they suggest cherts, are exposed in a few places, but the microscope shows that the quartz grains are clastic fragments. Lenses of tuff or graywacke have not been seen on the upland, but one thin lens of a light greenish-gray, fine-grained chloritic schist, 30 feet east of State route 346, where it crosses the western border of Bennington quadrangle, shows the same characteristic "pavement" of equant albite crystalloblasts with a fine-grained quatrz-sericite groundmass as was noted southeast of Center Berlin. This ledge may or may not be a part of the upland block.

Internal structure.—It is difficult to find a single ledge in this area in which the flat-lying thin quartzose layers are not folded. Small contortions, with a wave length of about ½ inch, are exceedingly common, and, though most of the axes trend north-northeast or northwest-southeast, other directions dominate in local areas. Near the summit of Mount Anthony, for instance, the axes of folds plunge south for the most part. The warped and folded layers are usually sliced through at short intervals by slip planes dipping at low angles east, and some exposures near the summit of Carpenter Hill are truly remarkable examples of shear folding, both on a megascopic and microscopic scale (Pl. 12, fig. 3). Where the axes trend north-northeast, the limbs are overturned to the west, but, where axes are oriented otherwise, the components of movement along the slip planes show considerable variation, and rarely is it possible to say in which direction a given mass of rock has been lengthened most.

Borders of the upland.—In this area, a thorough knowledge of the contact relations would probably solve more structural problems than anywhere else in the whole region. Unfortunately, only a few contacts are exposed; others permit of more than one interpretation, and still others are concealed.

The upland is bordered, on the east, by the limestone complex that underlies the valley of western Vermont (Pl. 1). On the north, the argillaceous rocks merge with those of the Taconic Range at Mount Anthony, and the limestone tract of North Pownal separates the upland, at least in its central latitude, from the quartz-chlorite schists of the Taconic Range proper. On the southeast, the quartz-chlorite schist of the upland extends toward the broad front of the Lower Cambrian Cheshire quartzite on the southwest slopes of the Dome, but the contact itself is buried under thick till. Nothing seems to have previously been published on the contacts of this mass of schist, and, as they are of special importance, they are described in some detail.

Eastern border: On the north side of Mount Anthony, the argillaceous rocks are underlain by two limestone tracts of Ordovician, probably Beekmantown, age. The slate that separates them on the northwest and north cannot be traced along the east slope of the mountain and is lost in slope debris. At the higher elevations, black slate and greenish-gray, glossy phyllitic rocks dip eastward and southeastward, conformably folded with each other as far as can be judged. The simple synclinal structure of the northwestern part of the mountain passes into an accordion structure of small folds whose axes still trend north-northeast, but the limbs are more compressed. Along the eastern base of Mount Anthony are limestone outcrops, and eventually slate and limestone meet in the bed of a small brook, about 300 feet southwest of the buildings of Everett Estate (Locality 27; the estate buildings are located at the little curve of the 880-foot contour, 1.1 miles due south of the letter "n" of the word, "Bennington Center", 15′ Bennington quadrangle). Black slate, dipping 30° SE., here rests conformably on well-bedded limestone, but, where the slope steepens, a northeast-striking high-angle fault interferes, and west of it is a continuous mass of black slate. This underlies the bed of the brook westward, up the east slope of Mount Anthony. Where it emerges at the crest, northeast of the summit, the black slate appears to rest on green quartz-chlorite schist. However, the dip of the rocks so nearly parallels the slope that one cannot be sure of the relationship between the two rocks. The exposures, as so often, are too low and discontinuous to decide the matter. Nevertheless, the writer believes there is no fault here. Instead, he assumes that the black slate was originally deposited on top of the green schist, and that both rocks have since been tilted and deformed (Pl. 1, cross section 1).

About 1000 feet south-southwest of the fault just mentioned, in the bed of another, parallel brook, slate and limestone have been brought to the same level by another fault, and again the magnitude of the displacement is unknown (Pl. 1, cross section 2).

Half a mile south of this fault exposure, the eastern border of the slate can be mapped fairly accurately. The slate ends abruptly along a series of precipitous cliffs in a pasture just north of the southernmost brook (shown on Bennington quadrangle) that drains the east slope of Mount Anthony. The structure within the argillaceous rocks is so confused

that not much can be said about their probable continuation, but 400 feet east, at lower levels, limestone dips steeply to the west and east. As the slate contact is followed farther south, it rises obliquely along the slope, and, a quarter of a mile south, the contact is 350 feet higher than at the east base of Mount Anthony.

The slate contact is concealed by talus for 1½ miles south. Dolomitic limestone, with thin sandy interbeds, crops out 100 to 200 feet east, but the true relations between the rocks remain unknown. In places, the black slate dips gently east and would run into the limestone if extended. However, the cliffs on the steep slope have suffered so much from creep that the dips may record slumping, and not the true dip of the rocks. If the argillaceous rocks lie horizontal, the limestone might be assumed to underlie them. Still another possibility should at least be mentioned. The rock adjacent to the limestone is a black, glossy, fine-grained slate, not the greenish chlorite-quartz schist that underlies most of the upland. The exact relationships between the two argillaceous rocks are not clear. As an alternative hypothesis, it could be assumed that the black slate rests on top of the limestone, and that a high-angle fault separates it from the chlorite-quartz schist. Exposures along the eastern boundary of the upland are poor, and the possibility of this relationship was recognized too late during the field work to be specially checked. However, in view of the peculiar exposures on the northwest flank of the Dome this matter deserves further study.

At the top of Carpenter Hill, the argillaceous rocks project slightly to the northeast. At the top ledges, greenish-gray, quartzose schists dip gently southeast, but, a few hundred feet north and northeast of the summit, the rocks are cut off by one or more faults, and limestone, with several thin lenses of black slate, is overlain conformably by black and dark-gray slate. This limestone may or may not underlie the top ledges of Carpenter Hill; at the critical places till conceals the bedrock.

About 200 feet southeast of the summit of Carpenter Hill, the argillaceous rocks dip about 10° W. and form a low cuesta, facing east. Ledges of limestone, directly east of it, and 3 feet below the apparent base of the siliceous schists, show flat *s* planes, but the actual superposition of the rocks is not exposed. Both the basal 3–5 feet of the schist and the underlying limestone for some 30 stratigraphic feet below the level of the contact display lineation in nearly horizontal northwest-southeast directions, and axes of folds strike in the same direction. Blocks of cream-colored dolomite have been ruptured, and, as in other places, the calcitic gray limestone has flowed around and between the fragments. In contrast with these intensely deformed and recrystallized rocks, the limestone farther down the east slope shows neither folding nor any other form of deformation. Details of bedding, sandy, even cross-bedded layers are seen in places, and it is apparently somewhere in the lower slopes of Carpenter Hill that Walcott (1888, p. 238, "three miles south of Bennington Center") found Ordovician fossils. Surely, if this mountain land is penetrated by flat thrusts, this exposure, south of the summit of Carpenter Hill, answers the requirements of the hypothesis. Yet we do not know how far the one rock has moved over the other, and nowhere else along the 7-mile-long contact have we seen quite the same relations.

South of this locality, the contact is concealed, and limestone and argillaceous rocks are 100–200 feet apart; strikes and dips on opposite sides are commonly so different that a fault seems to separate them. In a few places, a high-angle fault is necessary because the schist, if extended eastward, would run into the limestone. To restore the position of the rocks prior to faulting, the western block would have to be raised as in the two faults east of Mount Anthony.

A large scarp, north of Pownal Center, exposes beautifully the shear-folded, flat-lying layers of the chlorite-quartz schist. Northwest-trending lineation and, occasionally, axes of folds in the same direction suggest that the rock has been intensely drawn out northwest-southeast, but talus and till conceal the contact with the limestone.

Loc. 28. Three-fourths of a mile southeast of Pownal Center, the limestone-schist contact is exposed on the north-northeast slope of *hill 1440'*. Limestone underlies the schist on the north and northeast side and dips at about 45° S. and SW. As the beds are followed in this direction, up the hill slope, the limestone becomes tan-colored and pinkish, takes on thin films of quartz sand which become thicker and more micaceous upward, and at the top of an interbedded zone, 35 feet thick, the rock is the common chlorite-quartz schist. In this interbedded zone, the rocks show no signs of deformation; neither folds nor slip zones nor lineation were found. The schist is such a thin cap over the limestone that a shallow depression, just west of the crest of the hill, re-exposes the limestone. Several additional outcrops of limestone are seen near the top of the hill, but farther southeast the schist appears to be terminated by a high-angle fault so that the original relations between the two rocks are disturbed. The contact is concealed farther southeast.

WESTERN BORDER: The northwestern border of the Pownal upland is difficult to define. The quartz-chlorite schist of the Taconic Range that reaches Mount Anthony from the southwest is sufficiently similar to the more quartzose schists of the Pownal upland to make discrimination in the forested terrain extremely difficult and uncertain. The rocks of both areas are so folded and contorted that a structural separation is likewise unsafe. To draw an arbitrary fault, separating the upland rocks from those that build up Mount Anthony and the Taconic Range, would force the interpretation far beyond a safe basis of evidence.

The only, and unreliable, suggestion that a fault may be present in the critical zone could be based on one or two small limestone outcrops, the one in the bed of the south-draining brook, 3 miles north of North Pownal, the other one exposed in a culvert of a private road that reaches Mount Anthony from the southwest, at an approximate elevation of 2000 feet, about half a mile southwest of the summit (Pl. 1). It might be argued that the limestone represents the last remnant of a smeared-out carbonate horizon; however, the surrounding schist at Mount

Anthony shows no extreme deformation, and the first outcrop is surrounded by till, and perhaps not even in place. Therefore, we believe the possibility should be considered that both areas are actually parts of the same mass of rocks, still connected at the north, but separated farther south by the carbonate belt of North Pownal.

The western border northeast and east of North Pownal is concealed. Dips are gently to the east, and possibly the schist has been thrust westward over the limestone mass. The eastern one of the two limestone branches seems to rest on its west side on the quartz-chlorite schist, to judge from a few low outcrops on the northwest slope of hill 1040′, 1½ miles northeast of North Pownal (North Pownal quadrangle, N.Y.-Vt.).

Near the village of Pownal, the southwest base of the upland is underlain by black, fissile slate, and limestone lenses are interbedded with it. The association is so typical of the Ordovician rocks that Prindle and Knopf (1932, p. 288) regard both rocks as of that age. Several limestone lenses have been mapped (Pl. 1) between State route 346 and U. S. Highway 7, north-northwest of Pownal, but the boundaries of the black slate zone are largely concealed by till. At the western base of this series (hill 660′, just east of State route 346, at the west border of 15′ Bennington quadrangle, Vt.), the chlorite-quartz schist appears again, interbedded here with limestone lenses, about 10 feet thick, and it carries a thin lens of highly albitized ?tuff, referred to on a previous page. This rock has been marked (Pl. 1) by the designation, alb., and a thrust is drawn west of it. It may be bordered by another thrust on the east, separating it from the black slate series, but this is not drawn on the map because the relations could be explained otherwise.

The black slate is well exposed in cuts of U.S. Highway 7 and in old cuts of an obsolete trolley line that ran from Pownal to Bennington, a short distance west of the highway. The rocks there contain small biotite crystalloblasts, ⅛ mm. across, and larger ones of magnetite, but in the rocks southeast of Pownal no trace of metamorphism was found. The black slates can be followed eastward, up the hill slopes, to elevations of about 1100 feet (roughly coinciding with the albite isograd as shown on Pl. 1). Dips are gently to the east, but till conceals the contact with the greenish-gray, highly quartzose schists of the upland proper. Moreover, in the easternmost outcrops, the black slate is rather arenaceous, grayish, and one may not everywhere be sure where the boundary between the two groups of rocks should be drawn, for instance on the west slope of Mann Hill.

Loc. 29. One and a half miles southeast of Pownal, U.S. Highway 7 makes a curve convex to the northeast. Just south of it, Walcott (1888, p. 237) reported Ordovician gastropods in limestone ledges by the road. If one climbs the steep slope, just east of the highway curve, limestone ledges end about 200 feet above the level of the highway, and a dull greenish-gray quartz-chlorite schist overlies the limestone along a horizontal, conformable surface which is perfectly exposed for about 200 feet (Pl. 6, fig. 4). The contact oscillates somewhat and is offset slightly in places by high-angle faults that have dropped the west side a few feet. The argillaceous rock is a quartz-rich muscovite-chlorite schist and contains scattered crystalloblasts of albite (ab_{90}) (Pl. 12, fig. 2). In several ledges, directly above the limestone, pebbles of quartz are seen, ¼ to ½ inch across, a few even over 2 inches in diameter.

It is exceedingly difficult to interpret this contact. The rocks adhere to each other but are not interbedded, the limestone has been fractured and silicified in places, and in such zones the green schist has been broken, and fragments of it lie in warped position on the limestone, cemented to it. The lowermost 3–5 feet of the schist do not show particularly intense deformation, although a faint lineation, trending west-northwest, can be recognized on nearly every *s*-plane. The quartz pebbles seem to be true clastic fragments and have not been drawn out into ellipsoids, and in the limestone one looks in vain for signs of flowage, lineation, or other indications of severe mechanical deformation. Some 15 feet above the contact, the green schist is interbedded with black slate, or fine-grained phyllitic schist, and in the zone of interbedding were noted subhorizontal slip planes, spaced at 4- to 10-inch intervals, striated in east-west and west-northwest—east-southeast directions. Elsewhere small folds plunge gently north-northeast.

Unfortunately, the southeastern continuation of this contact is concealed. About ½ mile east, on the west bank of a south-flowing brook draining the east slope of Mason Hill (shown on Bennington quadrangle), one small ledge of a flat-lying, fissile black phyllitic slate is surrounded by till, and 100 feet east of it is a large scarp of the Cheshire quartzite. There is good

evidence that a high-angle fault separates the quartzite from the black slate; but whether this rock is continuous with the schist that overlies the limestone cannot be determined.

Till covers the southeast border of the upland. One small outcrop of black phyllitic slate, striking similar to the one just mentioned, was seen on the west side of the brook that drains the northeast side of Mason Hill, at an approximate elevation of 1450 feet (in the bottom of a former well that had collapsed through abrasion from the brook during a recent flood). It may, or may not, be a part of the upland rocks.

Before passing on to the last area, it may be well to summarize briefly the problems which the Pownal upland poses. It may be assumed that the quartzose schists overlie conformably the carbonate rocks south of Bennington, as they appear to do at the top of Carpenter Hill. As the carbonate rocks there lie flat, and have furnished Ordovician fossils, the overlying schist might be considered also of Ordovician age. However, if the evidence of thrusting in the ledge south of the summit of Carpenter Hill is considered, the age of the schists remains in doubt. Yet we have next to consider the interbedded contact, southeast of Pownal Center. If the limestone here is Ordovidian, the quartz-chlorite schist would move back to the Ordovician. However, according to Prindle and Knopf (1932, p. 272), there is probably dolomite of Lower Cambrian age also in the valley south of Bennington. As the limestone in question has not yielded fossils, its exact age is open to question; but if the limestone is Lower Cambrian, the schist of the upland would overlie rocks of various ages, and the contact, at least at one place, ought to be a disconformity or unconformity, a thrust, or an unconformable contact subsequently distorted by a thrust. We are just as much in the dark at the contact east of Highway 7 (Loc. 29), where the greenish schist overlies presumably Ordovician limestone along a flat surface which could be interpreted as a horizontal thrust, but also as a surface of superposition subsequently slightly dislocated. It is instructive to visit the thrust contacts at North Pownal and this contact on the same day; the wealth of detail in which the thrust at North Pownal is recorded certainly stands in striking contrast to the dearth of such features at Locality 29. Lastly, we have the till-covered stretch north of Pownal where it is so difficult to say where the green schists end and the black ?Ordovician slates begin. Until we know the true ages of the rocks in these particular outcrops, the nature of the contacts of the Pownal upland cannot be deciphered and admits of more than one interpretation.

6. Green Mountains

The Precambrian rocks that underlie most of the Dome (Bennington quadrangle) and adjacent territory in the Green Mountains consist of an oldest group of migmatites and injected and granitized amphibolites. The name, Mount Holly series, was applied by Whittle (1894) to rocks he studied farther north. Whether the mixed rocks near the Dome belong to this series or not we leave open. The gneisses and amphibolites exposed near Mount Holly, Vermont, over 30 miles to the northeast (Wallingford quadrangle), are not particularly similar to the migmatites, and much field work in the southern Green Mountains is needed before these rocks can be correlated.

Commonly, the gneisses display single, large crystals of orthoclase, which may be so crowded that the rock gives the impression of a coarse-grained granite, or granite gneiss, and this phase has been called the Stamford gneiss. Blue quartz is characteristic of this rock, and large, fresh exposures of it show a beautifully mottled pattern of crystals.

The Precambrian rocks are overlain unconformably by the Lower Cambrian Cheshire quartzite, in which micaceous zones have yielded fossils (Walcott, 1888, p. 236). The quartzite is a highly siliceous, white, yellow, or pink rock, generally fine-grained and thinly bedded, but locally conglomerates, arkoses, and muscovitic zones are seen.

The quartzite forms the summit of the Dome, and the unconformity at its base is repeatedly exposed northwest and west of the summit. However, lower on the northwest slope, the Precambrian rocks have been thrust obliquely up to the northwest and have overridden the quartzite (Pl. 1, cross section 6). The thrust zone can be mapped for about a mile, and in several places gneiss and quartzite approach to within

3 or 4 feet. All rocks exhibit the typical thrust structure with *a* lineation trending east-southeast, and plunging nearly parallel to the dip of the *s* planes. A mile and a half southwest of the summit of the Dome, the quartzite extends much farther west; it retains, however, the same highly deformed structure, with eastward-plunging lineation. On both slopes, the quartzite is abruptly terminated on the northwest by a high-angle fault which has dropped younger rocks to the level of the quartzite. The most important locality is a spur on the northwest slope of the Dome, Locality 30.

Loc. 30. The spur is plainly shown on the topographic quadrangle map. At the base, and for about 250 feet up, are pink limestones with thin sandy interbeds, dipping gently southwest. Slip cleavage dips east-southeast but is not everywhere in evidence. The pink limestone is succeeded by light-gray, partly massive carbonate rocks with folds here and there; a large fragment of quartzite embedded in the limestone is exposed by a road that climbs up this spur to a single cottage. Southeast of, and at the level of this cottage, the sequence continues with about 100 feet of a dark-gray, quartzose phyllite, well exposed in a series of cliffs southeast of the cottage. This phyllite, in turn, is capped by about 7 feet of limestone, lying substantially horizontal as does the underlying rock; and topping the section is another 10 feet of gray phyllitic slate. This entire pile of rocks is cut off by a high-angle fault that places the upper phyllite against Cheshire quartzite. The phyllitic rocks are thinly bedded and suggest undisturbed rocks, although under the microscope myriads of very small contortions of the muscovitic groundmass show that at least some of these layers have also been distorted. The groundmass of small quartz grains, and muscovite strands, with little chlorite, contain locally a few albite grains, about .02 mm. across, as well as brown tourmalines, ¼ to ½ mm. in length. Larger folds are absent, and lineation, trending southeast-northwest, is not pronounced. The southeast end of the spur is crossed by a second fault, trending northeast, and possibly the upper phyllitic slate is the same as the lower one, but placed at a higher position by this second high-angle fault. Exposures are not adequate to decide the matter.

The southwestern continuation of the northeast-trending high-angle fault is offset about a mile to the west. Whether this has been accomplished by a flexure or by the apparent horizontal component of a fault of northwesterly strike (Pl. 1) is hard to say. Such a fault might connect with the one at the east slope of hill 1440′, southeast of Pownal Center, but till covers the interval.

This completes our survey of the area. We now summarize briefly the results of the survey of the Taconic area, and discuss the possible interpretations of the structure.

Summary of Taconic Area

In its southern part, the Berlin valley is underlain by a belt of black slate and limestone, very probably of Ordovician age. On the whole, the slate is less deformed than the greenish argillaceous rocks that surround it on the west and east, and in several places evidence suggests that an underlying complex of greenish argillaceous rocks has been pushed up, into, and through the ?Ordovician rocks.

In contrast to the absence of thrusts on the west side of the Berlin valley, there is satisfactory evidence of a broad zone of thrusting, from the vicinity of Hancock, Massachusetts, northward at least to Center Berlin, and probably on beyond North Petersburg to the northwest slopes of Mount Anthony. Whereas the evidence for the thrust in the southern area is based on lithology, general stratigraphic relations, and structure, the evidence in the northern area is structural only.

An excellently exposed, eastward-dipping thrust borders the North Pownal limestone zone on the west side, and along the borders of the Pownal upland, also, thrusting is indicated in places. But elsewhere the evidence is ambiguous.

The Precambrian core of the Dome has been thrust northwestward into the Cheshire quartzite, producing the characteristic thrust structure in all rocks involved.

The flattest thrust planes are immediately west of the Green Mountain front. Thus, at the Dome, the Precambrian rocks rest on the quartzite at angles as low as 15°. If the entire contact of the Pownal upland is interpreted as the trace of a thrust sheet, that sheet would have a very gentle westerly slope (from elevations near 1300–1800 feet on the east, to about 800 or 900 feet on the west). The thrusts that emerge along the western and northwestern base of the Taconic Range, on the other hand, dip 30°–60° E., or steeper.

Nature of the Deformation

General Discussion

An attempt to interpret the observations presented meets with infinitely greater difficulties than it would in the central Appalachians. In this part of the Taconic Range, we lack as yet

a reliable stratigraphic section and cannot estimate, therefore, the displacements along faults. Although Keith (1932, p. 399–404) proposed a number of formations for the extreme northern termination of the range, near Brandon, Vermont, the nature and relations of these were apparently so obscure that Kaiser (1945, p. 1091) and Fowler (1950, p. 38) rejected them in favor of the older formation names of Dale (1899) and Cushing and Ruedemann (1914). No modern, detailed study is available for the 60 miles of the Taconic Range separating the present area from the Brandon district, and the stratigraphic and lithologic facies conditions along the west base of the Green Mountains have not been examined since the time of Pumpelly (1894). Obviously, therefore, an attempt to reconstruct some of the movements that affected this region is incomplete and will be replaced by better syntheses only as the rocks and the whole area are studied in detail.

The thrust zone that extends from the Hancock, Massachusetts, area northward to North Petersburg, and northeastward to Mount Anthony, plays somewhat the role of a tectonite front (Fellows, 1943, p. 1411, Pl. 2). East of the zone, *a* lineation is common; west of it, it is lacking or seen in sporadic outcrops only. As the writer has detected no trace of a low-angle thrust on the edges of the graywacke areas, he prefers to assume that these districts are underlain by essentially autochthonous formations, in which the competent graywacke layers have been folded, fractured, and considerably faulted, but neither been moved as a unit from their original site of deposition nor been deformed so intensely that a pervading east-southeast-trending *a* lineation has been developed in them.

All this is different east of the Berlin valley thrust zone. From its trace to the Precambrian core of the Green Mountains, lineation trending east-southeast pervades every cubic mile of rock. It can be seen on a microscopic scale in suitably oriented thin sections; it is recorded by thousands of spindle-shaped quartz-sericite pellets in the Taconic Range and by dozens of elongated pebbles, at Sugarloaf Hill and vicinity; it is seen in the threads and strands of epidote, west-northwest of Hancock, and in the tracts of graphite and chlorite in the limestone southeast of that village. The structure is extraordinarily strong where the Cheshire quartzite has been overridden by the Precambrian rocks at the Dome, and the migmatites are themselves striated and grooved by the *a* lineation near the thrust. Throughout the Taconic Range, the structure element appears on countless east-dipping bedding planes, slip planes, cleavage planes, and fracture surfaces. Where clean east-west surfaces are exposed, as on State Highway 2 near Petersburg Pass, or in the railroad cuts near North Pownal, one can measure small displacements along these planes and ascertain that the lineation represents the direction of maximum lengthening of the rock masses. The *a* lineation is so uniformly developed that it seems to indicate a general and, it seems, simultaneous stage of flowage, or protrusion, of the entire terrane between the Berlin valley thrust and the Green Mountain front, obliquely upward to the west-northwest. In the incompetent chloritoid-bearing quartz-chlorite schists and associated black phyllitic slates, the preceding asymmetric folding apparently had not reached isoclinal stages when the uniform flowage with *a* lineation took over.

In this belt of argillaceous rocks, about 8 miles wide, we know of no compelling evidence of a flat thrust. The innumerable exposed slip planes, with microscopic or larger displacements such as the one at Butternut Hill, or the 17 slip zones in the North Pownal railroad cut, dip 30° and 60°E. If a low-angle thrust is present, the writer has failed to find it and would be grateful if the omission were corrected. Only the thrust belt on the west slope of the Dome exhibits dip angles of 15° to 35°; its thrust planes may, as a whole, be flatter than those in the Taconic Range, but it would be desirable to have the few observations from this one, small area supplemented by others from a larger zone along the west border of the Green Mountains.

It could, nevertheless, be argued that the deformed rocks of the Taconic Range are an allochthonous thrust sheet, and that all the visible traces of powerful deformation were developed after this sheet occupied its present place. Admittedly it is impossible to disprove this argument, but it seems even more difficult to prove it. Many reasons have been listed previously (Balk, 1936, p. 767). An important one is the absence in the Taconic Range section

of anything resembling a strong stratigraphic horizon that could be imagined to have acted as the sole of the travelling sheet. Had we a Madison limestone, a basal dolomite-limestone complex like that of the Muddy Mountain or Glendale thrusts (Longwell, 1928, p. 109; 1949, p. 937), or a strong volcanic complex like the Excelsior formation in the Tonopah and Hawthorne districts, Nevada (Ferguson and Muller, 1949, p. 17), the mechanism of long-distance movement would be understandable. Though thrusts in the Basin and Range province commonly cut across formations and bedding planes, the fact remains that mechanically strong, competent rocks are nearly everywhere present to transmit the stress. Nearly all recent, detailed accounts of these remarkable thrusts emphasize considerable brecciation and fracturing of the rocks near the thrust surfaces. In the Taconic Range, nothing of the kind has been found. From the early researches of Hayes and Willis, to the most recent work by King and associates in the southern Appalachians, thick dolomites, limestones, or sandstones are reported at the base of most of the greater thrust sheets, and the nearly invisible thrusts where shale overlies shale (Neuman, 1951, p. 746) are local exceptions, not the rule, if the writer understands the literature of that area correctly. Moreover, blocks of limestone wrapped into the lubricating shale have been found (Wilson, 1935, p. 62). According to Neuman's sections (1951, Fig. 4, p. 751), the thrust blocks in the Great Smokies are composed, for the most part, of strong members of the Ocoee series, and only locally by a sericite phyllite of unknown age. The so-called Sandsuck shale includes, according to this author, "interbedded impure limestone, conglomerate, and sandstone" (1951, p. 744). Perhaps, where these interbeds are voluminous, that formation was fairly resistant.

Several thrusts in the Basin and Range province as well as in the southern Appalachians cut across bedding planes and have moved a complex of tilted and folded beds over an autochthonous series of rocks having contrasting structure. If a thrust in the Taconic area were of this nature, it undoubtedly would have been detected from contrasting strikes and dips above and below.

Deformation of the Pownal Upland

In the writer's opinion, the Pownal upland is the only area in which low-angle thrusting can be assumed. The evidence is not compelling, but, as exposures are poor, and the ages of the rocks uncertain, it may be advisable to point out some possible interpretations of its structure, and the evidence for and against each.

ASSUMPTION 1: The argillaceous rocks of the upland are a flat, allochthonous plate. The strongest evidence for this assumption is the contact just south of the top of Carpenter Hill, where the underlying limestone is intensely sheared in the west-northwest direction, shows abundant evidence of rupturing, accompanied by flowage of the more ductile calcite limestone, and where this deformed zone seems to pass downward into undeformed, horizontally bedded limestone. The *a* lineation in the lowermost 10 feet of the argillaceous rocks contrasts with less intense traces of deformation above. The contact east of U. S. Highway 7 (Loc. 29) could also be interpreted as favoring this interpretation, although the visible deformation in the argillaceous rocks is less intense, and the underlying limestone happens not to show much strain in the few hundred feet where it is exposed; perhaps, also, some deformation is masked by silicification. The interbedded contact of limestone and quartz-chlorite schist, southeast of Pownal Center (Loc. 28), must lie either below or above the thrust plane. If it lies below the thrust, we cannot point to any exposures structurally above the interbedded contact where such a thrust is suggested by structure, contrasting lithology on opposite sides, or any other features. In fact, it would be unusual, to say the least, that quartz-chlorite schist containing magnetite crystalloblasts of identical habits and sizes should crop out on opposite sides of the thrust (outcrops on the northwest slope of Mann Hill, and on hill 1440′ (Loc. 28), about 10 stratigraphic feet above the top of the interbedded limestone).

If the hypothetical thrust is placed below the interbedded contact of hill 1440′, we have the old problem of deriving a carbonate rock from the east slopes of the Green Mountains-Berkshire Hills. If any reliance is placed on the

direction of the *a* lineation, we must derive the allochthonous rocks from the east-southeast. Therefore, the only formation east of the Precambrian core of the Green Mountains would be the Conway (or Waits River) formation (Emerson, 1898, p. 188; Currier and Jahns, 1941). The writer is familiar with this rock from prolonged field work in northwestern Massachusetts and southern Vermont. The lithology of the Conway formation is so typical and different from any of the limestones west of the Green Mountains that it is extremely doubtful if such a correlation could be maintained; to the writer's knowledge, no such suggestion has ever been expressed in the literature.

One other interpretation could perhaps be tried. It might be assumed that the lowermost argillaceous beds, with quartz pebbles at Locality 29, are the same as the basal conglomerate that overlies the Precambrian rocks at the crest of Hoosac Mountain, 7 miles southeast of the Dome (Pumpelly, Wolff, and Dale, 1894, p. 55–59; 74–77; Prindle and Knopf, 1932, p. 287). The rocks may be Lower Cambrian as assumed by Prindle, though the possibility of an Ordovician age should not be discarded, in the writer's opinion, until the entire intervening area has been mapped in detail, and the results published. If the argillaceous rocks are considered Lower Cambrian, and the interbedded limestone section is included in the thrust sheet, it might be assumed that the limestone at the base belongs to the Lower Cambrian limestone series thought to be present in the general neighborhood of Bennington (Prindle and Knopf, 1932, p. 272). If the argillaceous phases overlapped in places upon the crest of the Hoosac Range, perhaps some of this limestone reached as far to the east. Under these assumptions, the Pownal upland could probably be explained as a thrust sheet having moved approximately 7 miles to the west-northwest. It still would be necessary to gather some evidence that the underlying, interbedded limestone is separated from the rest of the limestone terrane by a thrust, but the extensive till might be assumed to cover the critical zone. There are several additional objections to this derivation of the rocks, but the purpose here is to picture the main problems that the reconstruction of the structural history must try to solve.

ASSUMPTION 2: The argillaceous rocks of the Pownal upland are a plate of autochthonous Ordovician rocks, subsequently deformed in places by low-angle thrusting with moderate displacements. The interbedded contact at Locality 28 would support this interpretation for the limestone may well be Ordovician. Lithologically, the quartz-chlorite schist differs from the habitually black or dark-gray slate that accompanies the limestone tracts in the region in being much more quartzose and lacking the graphitic dust. This may not be too serious a matter because the Normanskill and Snake Hill formations on the west side of the Taconic Range are commonly cherty, or have sandy and gritty intercalations (Ruedemann, 1930, p. 97, 118; Fowler, 1950, p. 35; Dale, 1904a, p. 31–36). The sheared contact south of the top of Carpenter Hill could mean that this particular mass of rocks has been compressed and sheared off its base an unknown but perhaps small distance. The relatively weak deformation at Locality 29 would be satisfactorily explained. The thin black slate that appears at the limestone contact between Mount Anthony and Carpenter Hill might be a basal carbonaceous horizon between the quartz-chlorite schist and the top of the limestone. Possibly the small masses of black phyllitic slate, at the top of the section on the northwest slope of the Dome (Loc. 30), would be remnants of the same zone, and the originally overlying quartz-chlorite schist could have been eroded from these more highly elevated portions of the terrane. On its west side, the quartz-chlorite schist could overlie the black slate of the Pownal-North Pownal area, and the original surface of superposition might have been disturbed in places by thrusting as on the east side.

It is nearly impossible to find a contact on the north and northwest side. If the plate had been moved far, it might be presumed that the northwestern front of the mass would show appreciable crushing and shearing phenomena. They seem to be lacking although again the till cover may conceal the evidence. It seems,

nevertheless, that a deformed zone would have been found somewhere southwest of Mount Anthony where exposures are fairly continuous. Perhaps, then, it could be imagined that the argillaceous sheet was compressed from the east-southeast; that on that side the rocks were locally sheared off their base; but that the intense shear folding and small-scale crumpling of the sheet took up enough of the lateral compression to leave the northwestern portion essentially in place. Perhaps this picture is adequate to explain most of the observable features of the Pownal upland.

ASSUMPTION 3: The Pownal upland is an autochthonous mass of Lower Cambrian rocks, perhaps incorporating along its margins lithologically similar Ordovician sediments, the whole deformed in places by low-angle thrusts of moderate displacements as in assumption 2. The limestone at Locality 28 might be Lower Cambrian, or Ordovician. At Carpenter Hill, the contact would be a thrust where Lower Cambrian quartz-chlorite schist overlies Ordovician limestone, provided that the fossil locality referred to by Walcott (1888, p. 238) is not separated by a fault from the section that underlies the quartz-chlorite schist at the summit. The contact at Locality 29 would probably be a thrust. The place where Walcott obtained his Ordovician fossils, only a few thousand feet south of the contact, is easily identified, the limestone lies flat, and several outcrops connect the two exposures.

This third interpretation calls for some comments on the facies distribution in the early Paleozoic area of deposition. The easternmost outcrops of the quartz-chlorite schist are at present ¾ mile west of the Cheshire quartzite. In view of the exposed thrust at the Dome, one could move the site of quartz-sand deposition a little farther east—say, half a mile. The distance cannot be very great because the quartzite rests on the Precambrian core of the Green Mountains. Even the increased distance, a little over a mile, may seem too small to separate two obviously different sedimentary facies. However, before this correlation is dismissed, two points should be borne in mind. In the first place, the exact stratigraphic sequence and thickness of the Cheshire quartzite are very imperfectly known. No detailed study has been published since Walcott's (1888) and Pumpelly's (1894) investigations. Perhaps there is nowhere a complete section, but micaceous zones and lenses in the quartzite are seen at the Dome, at Clarksburg Mountain, 4 miles to the south, and it was from these that Walcott obtained Lower Cambrian fossils, east of Bennington. Whether the exposed thicknesses represent original dimensions, or sections distorted and thickened by isoclinal folding and thrust faulting, is largely unknown. Nevertheless, a section of quartz-muscovite schist, about 250 feet thick, interbedded with thin beds of quartzite, is exposed at the two summits, 1800′, ¾ mile east of Mason Hill (15′ Bennington quadrangle, Vt.). This is about 1 mile from the nearest outcrops of quartz-chlorite schist.

This leads to the second point. How far apart can a pure quartz sand (the Cheshire quartzite) and a quartzose mud (the quartz-chlorite schist) be deposited? Could it be assumed that the quartz-chlorite schist of the Taconic-Pownal upland was originally interfingered with the Cheshire quartzite, and that the subsequent deformation has merely disrupted most of the critical transition zone? This question cannot be fully answered at the present state of our knowledge, but a few facts from other geosynclinal areas may be worth comparing. In the Santa Maria district, California, Woodring and Bramlette (1950, p. 26-29, Pl. 2, cross sections E-E′, F-F′) describe the late Miocene-Pliocene Tinaquaic sandstone, a near-shore facies of a thick mudstone (Sisquoc) formation. In a measured section, the sandstone is 1450 feet thick, but it lenses out within a few miles from the former shore. The Gasworks sandstone of the Welsh Llandovery dwindles from 2000 feet to nothing within 4 miles; and the upper Llandovery (Clinton) muddy sandstone, at Llandovery, is 1500 feet thick but 2 miles away measures no more than 40 feet, (Jones, 1938, p. lxxxiii). According to Schwan (1949, p. 52, 57, and confirming letter to the writer, dated March 6, 1951), the late Devonian-early Mississippian Acker quartzite of the Harz Mountains is 180 miles long, 2–3 miles wide, and 450 feet thick. On its southeast side, this rock passes into graywackes appreciably thicker than the

quartzite, and on the northwest this rock passes within a very short distance into 30 feet of cephalopod limestone.

These randomly selected examples may be extreme, but the figures show that the dimensions with which we are here concerned are not beyond the possible. With Prindle, the writer believes that appreciable parts of the Taconic Range are made up of Lower Cambrian argillaceous rocks, and it seems quite probable that these rocks were interfingered with the Cheshire quartzite. The environment in which these early Paleozoic rocks may have been deposited is briefly sketched, even though the reconstruction is imperfect and incomplete.

We would begin with a trough, extending from the west slope of the Green Mountain area roughly to the Hudson, a distance of about 30 miles. This trough would coincide essentially with the one assumed by Ruedemann (1930, p. 26–28, 131), who envisaged a barrier separating the trough from another, larger basin farther west. This barrier cannot now be seen, but it may be recalled that the Precambrian basement rises through the Paleozoic cover, 30 miles south-southwest of the Austerlitz graywacke area, at Stissing Mountain (Balk, 1936, p. 761, 762). Possibly, the uplifted block is a fragment of a western barrier that enclosed the basin in early Paleozoic time. The basin, then, with a width of about 30 miles and an unknown but probably considerable north-south length, may be assumed to have received the early Cambrian series of argillaceous-arenaceous sediments while the Cheshire quartzite fringed the eastern and southeastern shores as a narrow beach deposit. The muddy bottom areas seem to have received ample silt and sand as evidenced by the numerous thin quartzites with chlorite cement and the abundance of siltstones in the series. If the interpretations of Kuenen (1950, p. 238–248), Kuenen and Migliorini (1950), and Pettijohn (1950) are applied, it may be assumed that a considerable part of the graywacke was brought to the floor of the basin through turbidity currents, and it is interesting that an appreciable amount of this debris appears to have come from the west, suggesting that the barrier on that side may have been mobile and unstable. Despite Shepard's (1948, p. 150) warning that the distribution of grain size does not necessarily give a true picture of the direction of the source of a sediment, the writer believes that the striking concentration of coarse phases of the graywacke, the conglomerate lenses, and the fact that individual beds of graywacke are thickest along the western border would be hard to understand if this sediment had come from the east. The numerous graywacke lenses in the Taconic Range, however, are all fine-grained, thin, and discontinuous as one might expect the toes, or distal endings, of more extensive sheets of graywacke to be.

Coarse, micaceous sand, silt, muds of various colors and degrees of fineness, as well as a few calcareous lenses are thought to have been deposited in irregular succession and distribution, and we are perhaps justified in picturing the conditions in this basin as somewhat similar to those described by Umbgrove (1949, p. 5–11, Figs. 8, 11) from the Indonesian sea floor and by Revelle and Shepard (1939, p. 252–258) from southern California, where an irregular bottom relief, combined with strong currents but seasonally varying directions of the wind, is partly responsible for considerable variety of sedimentary material, distributed in nearly parallel zones or, in California, more haphazardly. The environment in which the early Paleozoic argillaceous rocks of the Ardennes and the massif of Brabant, Belgium (Fourmarier, 1931, p. B365; 1934, p. 12–21; Kaisin, 1936, p. 115–121), and the Devonian shales, sandstones, and limestones of the Lahn basin (Paeckelmann, 1930, p. 596, 597, Figs. 4, 5; 1934, p. 237–245), western Germany, were laid down seems to illustrate similar conditions and invites comparison. The striking facies variations over short distances as described by Schmidt (1925, p. 227–231) and Weber (1934, p. 538–558) from the Devonian sub-basins south of the Ruhr district may well find analogues in the Lower Cambrian and Ordovician rocks of the Taconic areas. The association of siltstones, micaceous quartzites, and graywackes with greenstone lenses, tuffaceous zones, pillow lavas, or basalt is, of course, very common in geosynclinal troughs. Even the limestone lenses along the eastern border of the graywacke upland may be re-

lated genetically to the tuff lenses. Examples of such association have been reported from numerous areas in the western Cordillera, for instance from southeastern Alaska (Buddington and Chapin, 1929, p. 46).

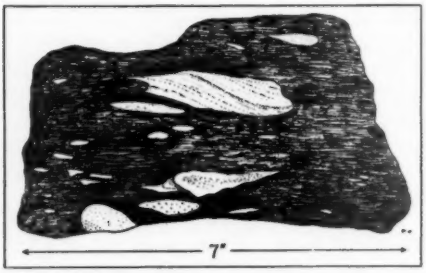


FIGURE 19.—BEDDING PLANE OF GRAY SLATE, ENCLOSING FRAGMENTS OF FINE-GRAINED MICACEOUS SANDSTONE

Fragments elongated parallel to intense lineation. Irregular shapes and curved bedding of several fragments suggest plastic deformation completed perhaps prior to solidification of sandstone. Northeast end of Breese Hollow, North Pownal quadrangle, N. Y.-Vt.

Some time before the Ordovician, this trough with its Lower Cambrian fill appears to have risen. Whether deposits of Middle or Upper Cambrian, or both have ever been deposited remains unknown. They are not now found in this area, but upon, and among, the arenaceous-argillaceous rocks of the Lower Cambrian series we now find black and gray slates, as well as limestones, of Ordovician age. The relationships have been studied by Dale (1899, p. 178; 1904a, p. 43; 1904b, p. 189). As he clearly recognized (Dale, 1899, p. 297), it is most difficult to reconstruct the original relations between these rocks because the Ordovician slates resemble the older ones, fossils are very scarce, and the ubiquitous cleavage commonly masks bedding planes, original contacts, or other separation planes. Nevertheless, Dale's conception of tracts of Ordovician sediments, deposited rather irregularly over the bottom sediment of Lower Cambrian age, seems competent to explain more observed features than any other concept.

Perhaps there is a slight unconformity between the Lower Cambrian and Ordovician rocks, at least in places. Several localities have been described where the structure of the ?Ordovician black slate and limestone differs from that of the surrounding greenish-gray, siliceous slates. The peculiar graywacke dikes, mentioned before, may also record movements of the trough prior to the deposition of the Ordovician rocks. It is impossible to say how important they have been, but it is common to find larger earth movements preceded by smaller ones. In this connection, we may mention a few observations that seem to illustrate peculiar kinds of deformation in the black ?Ordovician slate and limestone, not unlike submarine slumping as described by Jones (1939) from Wales.

Loc. 31. A road cut, west of the Rutland railroad track at the northeast end of Breese Hollow (North Pownal quadrangle, N.Y.-Vt.), exposes about 25 feet of gray and black slate, resting on impure, highly quartzose limestone beds. In contrast to the surrounding slate of the Indian Hill district, this small mass of slate is pervaded by extraordinarily intense lineation trending west-northwest on the nearly horizontal bedding planes. When first seen, this exposure was thought to mark the trace of an important flat thrust fault. However, several beds are replete with white blebs and pebbles of a micaceous sandstone whose major body axes lie parallel to the lineation as if they had been deformed when plastic. A few of the pebbles show internal stratification at an angle to the major elongation and have outlines so irregular that it is hard to believe that these pebbles were thoroughly cemented when they were aligned (Figs. 19, 20). They suggest that this mass of shale may have slipped on an insecure floor; that it contained numerous pellets and lumps of soft sand; and that these were drawn out during the movement and solidified afterward. A few hundred feet away, a thick lens of black cherty sediment shows equally strong lineation, but dipping as much as 70°. The rocks are extremely brittle and, had they been deformed after consolidation, they would probably have fractured into hundreds of minute bits. Yet the hillside is strewn with long slabs of the rock, essentially unfractured, and the slabs are elongated parallel to the lineation. One is again left with the suspicion that this rock was not yet solidified when it acquired its deformation structure. Under the microscope, the cherty layers are extremely fine-grained silty sand in which minute clastic quartz grains are embedded in graphitic films and sericite flakes. Perhaps this deposit was something like a lens of quicksand in the fine mud. The writer does not pretent to know the true manner in which these rocks were deformed but believes observations of this kind should be published. Only by patiently accumulating facts and observations on these poorly known rocks can we hope to learn something about their origin and the conditions under which they were deformed.

In line with this tentative interpretation is the evidence of intraformational slumping and folding in the ?Ordovician limestone, ½ mile south of

FIGURE 20.—VIEW OF BEDDING PLANE OF SLATE WITH SANDSTONE FRAGMENTS
From same locality as Figure 19, showing lineation and shapes of elongated sandstone pellets.

North Petersburg, already mentioned. It is not the only example of its kind. At the western base of Mount Anthony, as well as in outcrops on hill 1060′, 1 mile northeast of North Pownal (North Pownal quadrangle, N.Y.-Vt.), very similar small-scale contortions and folds of individual beds can be seen amidst flat-lying, apparently undeformed layers of dolomitic limestone with ample quartz sand that brings out the minute details of the contorted primary stratification.

Some time after the Ordovician rocks had been deposited, the area was deformed. The argillaceous rocks were folded, overturned to the west, and sliced through by millions of slip cleavage planes as well as by numerous thrust zones inclined to the east. The dominantly argillaceous rocks between the west front of the Green Mountains and the eastern border of the graywacke areas record a direction of greatest elongation obliquely up to the west-northwest. Where the rocks were strengthened by the graywacke terranes, folds seem to have attained greater dimensions, and steep dips are the rule. It is assumed that the folded strong rocks were broken by many eastward-dipping thrust faults, probably with moderate to small displacements, but owing to the poor exposures one cannot be sure of it. Low-angle thrusts, at any rate, do not seem to penetrate these rocks, and no known evidence suggests that they rest on foreign ground.

Conceivably, some of the Lower Cambrian rocks stood for a while as a group of islands in a shallow Ordovician sea, so that near them dominantly argillaceous Ordovician sediment was deposited—much of it perhaps derived from reworked older mud of Lower Cambrian shoals or islands. The ensuing folding and thrusting of all rocks would tend to obliterate these contact relations, but not necessarily the spatial distribution of the rocks. Thus a larger mass of Lower Cambrian rocks could be brought adjacent to a fringe of Ordovician shale, or slate. The contacts, marred by subsequent deformation, the lithologic similarity of the rocks, cleavage, and obscured by till could be overlooked even if they were exposed. If these Ordovician slate zones were narrow to begin with, they may have been cut out by subsequent thrusting on the west sides of a rising block, while, on the east sides of an uplifted block, there would be normal faults along which Ordovician and Lower Cambrian rocks might form a complex mosaic of fault slivers.

Some such conditions may help to explain the peculiar contacts of the Pownal upland. We know that in at least two places the surrounding limestone is Ordovician, yet the quartz-chlorite schist of the core of the upland may be, in part, Lower Cambrian; and the evidence is strong that it is flanked by Ordovician slate and limestone on the southwest. Perhaps the black slate of the eastern border, southeast of Mount Anthony, is also Ordovician, and the same could tentatively be assumed for the 100 feet of dark-gray phyllitic slate at the top of the limestone on the northwest slope of the Dome.

Concluding Remarks

It is unsafe to go further. Perhaps most, or all, of the conclusions or suggestions made here will be rejected in favor of others that can explain more facts better. In this region of dense underbrush, extensive forests, widespread till cover, low relief, and few fossil localities, facts and observations are hard to accumulate, and their interpretation is beset with difficulties. The structural investigation of the Taconic region is only in its beginning stages, and the writer is in full agreement with Prindle and Knopf (1932, p. 299, 302) that the interpretation of one particular area, to be satisfying,. should allow correlations with adjoining areas. The writer is not familiar with the Brandon, Vermont, district, but from a study of the literature he has gained the impression that the allochthonous nature of the Lower Cambrian rocks there is far from proven, and that the lithologic problem—that is, the contrast between the calcareous and argillaceous rocks—lies mostly within the admittedly autochthonous rocks, namely, the Ordovician Hortonville slate, or phyllite (Cady, 1945, p. 558), and the pre-Hortonville calcareous rocks. Perhaps the picture sketched in the preceding pages of a series of islands, exposing argillaceous Lower Cambrian rocks, with Ordovician mud laid down in their vicinity, as Dale (1904b) conceived it, may be competent to explain the present position of the severely disturbed rocks. At least further studies of this

area seem desirable before this concept is discarded.

Many important problems of the Taconic structure must remain obscure until other, related questions have been investigated more thoroughly. We do not know the structure of the crest of the Green Mountains-Berkshire Hills in sufficient detail to judge what deformation this important area has suffered. The same is true for the vicinity of Mount Greylock, despite Dale's excellent study of it (Pumpelly, Wolff, and Dale, 1894, p. 131–203). Isograds of metamorphic minerals have not yet been mapped for the area south of Hancock to the Millbrook quadrangle, New York (Fig. 1), and, even if Lower Cambrian rocks are voluminous in the Taconic area of this report, we still do not know how far south these ancient rocks may extend.

The source and nature of the weak metamorphism is not clear. It is interesting, however, to compare it with the very similar manifestations of metamorphism in the Cambro-Ordovician massifs of Belgium. In the core of the Ardennes, as well as in the massif of Brabant, late Cambrian, Ordovician, and Silurian slates develop chloritoid on a large scale (Corin, 1930, 1931). Magnetite crystalloblasts are also common, but the metamorphism has gone farther than in the present area, and garnet, andalusite, staurolite, and hornblende have formed in the vicinity of sporadic thin, aplitic veins. Small intrusives are also known. Considerable field work in the Taconic Range is needed for a better comparison of these two regions, but here is a rewarding field for future research in another of the many unsolved problems of the Taconic Range.

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Short Notes

EXPLORATION OF THE NORTHWEST ATLANTIC MID-OCEAN CANYON*

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It has been known for nearly a century that submarine canyons indent the continental slopes of the world. Although some early workers indicated canyons extending far out into the ocean basin, their data were so sketchy that it was generally accepted that the canyons ended at or near the base of the continental slopes. In recent years exploration has revealed that the canyons do not end at the base of the continental slope but continue across the continental rise to the abyssal plains of the ocean floor. Studies of sediments from the floors, walls, and seaward extremities of these canyons (Ericson, Ewing, and Heezen, 1952) have proved that powerful turbidity currents have repeatedly carried large volumes of sediment through the canyons and deposited them in well-sorted beds on the abyssal plains.

Until 1949, when Heezen, aboard the ATLANTIS, found four canyonlike depressions in the abyssal plain east of Newfoundland, there was no clue that canyons extended across the abyssal plains. It was not possible to alter the ship's course to investigate these depressions. It was considered that three of them (II, III, IV of Fig. 1) were crossings of a continuous feature which probably extended to form the other two (I and VI). The first opportunity to investigate this depression further came in 1952 when about 80 more echo-sounding traverses and 4 more cores were obtained in and near the depression, by the tug KEVIN MORAN and the ATLANTIS. Our conclusion is that the depression is a submarine canyon extending the full length of the Newfoundland and Labrador basins and well into the North America Basin. It has a known length of 1200 nautical miles and is probably longer than 2500 miles. It has a streamlike longitudinal profile, and the sediment cores imply that turbidity currents have been active in this canyon. It now seems likely that the only limits on the length or maximum depth for submarine canyons are those set by the length and maximum depth of the ocean basins in which they are found.

The canyon has been traced from 52°N. to 38°N. Latitude, a distance of 1200 miles. Over most of its known length it is box-shaped with a flat floor and steep walls. Although it is virtually impossible to measure the slope of the canyon walls with the echo sounder, it is probable that the slopes usually exceed 20° and in some cases may be vertical. Its depth below the adjacent deep sea floor ranges from 100 fathoms near Lat. 52°N., where the sea-floor depth is 2240 fathoms, to 10 fathoms near Lat. 38°N., and its width ranges from 3 to 5 miles except near the southwest corner of the surveyed area where in 2750 fathoms the canyon broadens and loses its boxlike shape.

The floor of the canyon slopes to the south from Profile I (Fig. 1) at 52°N. to Profile VI at 42°N. with a gradient of approximately 2.5 feet per mile (.045 per cent grade). In the region of profiles I through VI the basin floor slopes to the east from the east wall of the canyon for approximately 50 miles to the Mid-Atlantic Ridge with a gradient of 5 feet per mile. To the west of the canyon the ocean floor has practically no east-west gradient until it rises to form the continent. The east wall of the canyon is consistently lower than the west wall

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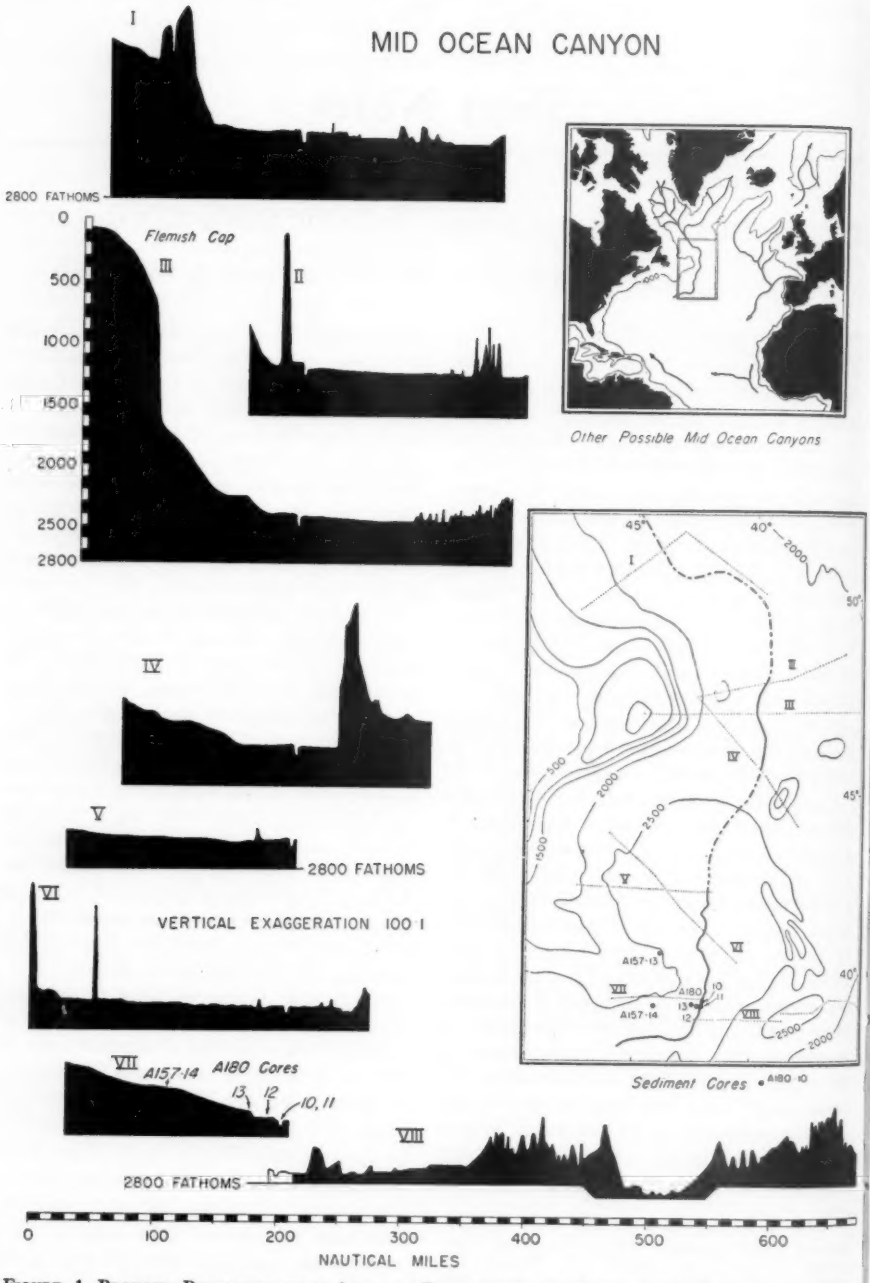

FIGURE 1. PROFILES REPLOTTED FROM ATLANTIS FATHOGRAMS AND BATHYMETRIC SKETCH AFTER TOLSTOY (1951) AND H. O. CHART 0955 OF THE NORTHWEST ATLANTIC MID-OCEAN CANYON

by approximately 10 fathoms. This difference is too great to be accounted for by a west-east slope of the bottom. After traversing the floors of the Labrador and Newfoundland basins the canyon runs through a rather narrow gap in the Southeast Newfoundland Ridge. This gap is not more than 30 miles wide and is probably much narrower. The canyon widens to about 5 miles in the gap and shallows to about 20 fathoms, apparently maintaining its cross-sectional area, and although its banks still maintain a southerly gradient of about 2.5 feet per mile its floor remains at the same depth (2720 fathoms) for about 50 miles in the gap before resuming downstream gradient. In passing through the gap the course of the canyon changes from essentially south to southwest by west, and this trend is maintained until the canyon widens, shoals, and loses its boxlike shape near Long. 44°W. The survey vessels had a tight schedule, and the study of the canyon had to be terminated when the trend became westerly.

Although the canyon showed definite signs of petering out on the last traverse it may well continue for many hundreds of miles beyond this point.

In one of two cores (180-10) from the canyon bed graded layers make up three-quarters of the 2-meter thickness of sediment recovered. In this section particle sizes do not exceed silt grade. Unfortunately the lower part, 2½ meters long, was lost by flowing through the core catcher. That this part was very probably sand is shown by traces of sand found on the core catcher and inside the coring tube. The upper 123-cm. section of the other core (180-11) from the canyon bed is composed of clay and silty clay in part as graded layers. This overlies 2½ meters of well-sorted fine sand.

A third core (180-12) was taken on the nearly flat western bank, 90 meters or about 50 fathoms above the canyon bed. Numerous graded layers show that fully three-quarters of the 4.6-meter total thickness was deposited by turbidity currents. Since fine sand makes up only 3 per cent of the total thickness we infer that the sediment was carried in the highest part of the turbidity currents.

A core from a point 18 miles west of the canyon contains a well-defined depositional break 45 cm. below the top. Although the foraminiferal clay below the break is no older than Pleistocene it is apparent from the faunal sequence that a part of the late Pleistocene section is missing in consequence of slumping or erosion. Two cores (157-13, 157-14) from the flanks of the Southeast Newfoundland Ridge contain similar evidence of slumping or submarine erosion.

In general the cores indicate that turbidity currents depositing sand and silt in the canyon feathered out on the banks. The burial of these sands and silts by over a meter of clay and silty clay would indicate that the last major turbidity current probably occurred in Wisconsin time. The extension of the canyon northward from the part in the rectangle in Figure 1 is inferred from the general topography of the area as shown in Smith, Soule, and Mosby (1937). It is considered highly probable that even a few continuous echo-sounding profiles across the two northern lobes of the Labrador Basin would confirm this hypothesis. It is highly probable that many other mid-ocean canyons will be found. Several strong probabilities in the North Atlantic are shown in Figure 1. Dietz (1953) has recently described depressions in the floor of the Indian Ocean which almost certainly represent similar canyons.

The fact that the canyon runs for hundreds of miles along the axes of the ocean basins, being limited in length and maximum depth only by the dimensions of the ocean basins, is difficult to explain under any of the various theories of canyon origin except turbidity currents and faulting. Faulting offers no explanation of the sediment relations or the streamlike longitudinal profile so easily explained by turbidity currents. It is probable that the mid-ocean canyon heads in the shelf canyons of Greenland and Labrador. Owing to freezing ice, the shelf waters of this region are seasonally of high salinity and low temperature, and are therefore favorable to the initiation of turbidity currents (particularly during the Plesitocene). It is considered probable that the Newfoundland, Labrador, and North America basins and any smaller depressions which might have originally divided them were initially filled to their respective sill depths by ponded sedi-

ment carried into the basins by turbidity currents travelling through the continental slope canyons of Greenland, Labrador, and the Grand Banks. The evolution of submarine topography then probably followed much the same pattern as the desert geomorphic cycle. That is, the highest basin trapped the sediments until they overflowed over the lowest outlet into a lower basin. The finding of an outlet permitted the establishment of a localized channel (mid-ocean canyon), and the downcutting of this channel as the outlet sill was lowered by erosion. As this process was repeated in each successive basin, the continental shelf canyons of each basin became tributaries of the mid-ocean canyon. The process now has presumably reached the stage where the Southeast Newfoundland Ridge has been breached, causing the channel to incise into the constructional floor of the Newfoundland Basin and transport sediment through it to the deeper basins to the south.

Once the canyon became established it began to act as a barrier to turbidity currents moving on the floor of the basin. This barrier action offers a possible explanation of the greater depth of the eastern bank of the canyon and the west-to-east gradient of the basin floor. The canyon originally ran down the deepest axis of the basins, cutting off the eastern part of the basins from flows down the continental slopes from the Grand Banks, and limiting sedimentation on the east side to flows from the Mid-Atlantic Ridge, the material settling out of the water and the feather edges of the turbidity currents overlapping the banks of the canyon. If natural levees were built along the sides of the canyon at any stage in its history they would have aided in holding back the flows from the west and in protecting the east side of the basin. The Mid-Atlantic Ridge, having many closed intermontain basins and a smaller area, sheds much less sediment than the continental margins.

Another possible hypothesis for the eastward gradient east of Flemish Cap (Profiles II and III) involves warping of the sea bottom. Under this hypothesis, at some time after the canyon had incised itself in the center of the basin, perhaps during the mid-Pleistocene, the west side of the basin rose some 100 fathoms, causing a tilt of the basin floor. With this explanation of the west-east gradient we must still appeal to the barrier action of the canyon to explain the difference in height of the two banks. Whatever the explanation, if the canyon was formed by submarine currents, it was formed before the present west-east gradient existed.

The relationship of the canyon to the Southeast Newfoundland Ridge may be more complicated than a simple sill and overflow situation. Evidence of interrupted depositional sequences on the ridge, which has been cited above, indicates that this ridge may have been tectonically active during the Pleistocene. In this case the mid-ocean canyon may have cut into the Southeast Newfoundland Ridge as it was warped up, and so in this region it might be regarded as a superimposed stream. Indeed, the characteristics of the canyon and surrounding topography in this region have the aspect of a mountain pass. However, echo-sounding traverses in this region are too few to define the situation clearly.

The knowledge of the mid-ocean canyon thus permits the description of a very important process in the topographic evolution of the deep sea basins—namely, "alluviation" by turbidity currents.

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